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ON CERTAIN RESULTS AND PROSPECTS OF TECTONOPHYSICAL INVESTIGATIONS[1]

by

V. V. Belousov

Tectonophysics, as understood by Soviet ecialists, is that science whose problem is e investigation of the physical mechanism tectonic deformations. Foreign scientists e the term in a broader and less definite nse, considering tectonophysics as the plication not only of physical but of various ner geophysical data to the solution of geo- ctonic problems of all kinds. They include tectonophysics, for example, the construc- n of general geotectonic hypotheses based t only on geological considerations but on ophysical considerations as well. We shall strict the following discussion to our con- pt of the content of tectonophysics and its oblems.

The phenomena of the crumpling of layers the earth's crust into folds and the forma- n of fractures, with subsequent movement parts of the crust along these fractures lative to each other, are of prime interest m both the theoretical and practical points view.

Long ago there developed a conviction that ologic methods alone are insufficient for e study of tectonic deformation and that ologic data may at least be substantially gmented by the application of other, physi- l methods. A geologist can observe only e final, immobile result of the process of formation of the rocks of the earth's crust. mparing different attitudes of rocks, he is le to make conclusions as to the stage of e process of tectonic deformation and as the conditions affecting its course and the al result. These conclusions, however, ll always have a hypothetical character, d many phases of the process will not be derstood or will pass unnoticed because e geologist cannot observe the process of formation itself. From this sprang the natural desire to reproduce tectonic defor- mations on models in order to observe and study those processes which are impossible to observe in nature. From this also, came the need to use the multitude of data on the conditions and laws of deformation of solids, in general, which have been accumulated in various branches of physics as a result of observations not on rocks but on other materials.

The first attempts at experimental repro- duction of the process of folding of strata are more than a century and a half old. They have been repeated many times during this period. Attempts have been made at experi- mental investigation of the mechanism of tectonic dislocations and of several other processes (cleavage, boudinage, etc.). Dur- ing the second half of the last century, a need developed to evolve a generalized theory of the physical side of tectonic deformation. This problem occupied such noted tectonists of that time as A. Daubree, E. Suess, A. Heim, E. Reyer, and B. Willis. B. Willis tried to reproduce experimentally some of the characteristics of the Appalachian folded structures and his experiments became classical.

At the end of the nineteenth century, G. Becker proposed a physical theory of forma- tion of tectonic dislocations which enjoyed a long success until it was proved erroneous. Later, attempts to work out certain phases of the physical theory of tectonic deformation were made by Mead, Bucher, Leith, and others.

In the first and second decades of our century, B. Sander developed an original and extremely complex science of microscopic deformation occurring in individual mineral grains during the deformation of the rock mass as a whole and devised microscopic techniques for studying them. Hans Cloos placed observations on fracturing of rocks at the center of his system of studying physical conditions of development of tectonic deformation. These works are generally

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[1] Paper read at the General Assembly of the ction of Geological and Geographical Sciences, ademy of Sciences, U.S.S.R., June 17, 1958.

known and do not require reference.

It must be said that this long period, which we may regard as preparatory, did not contribute a great deal toward the development of scientific tectonophysics. This was the result of, in many cases, the use of incorrect methods. In setting up tectonic experiments the investigators commonly tried to reproduce only the geometric similarity to structural forms and completely disregarded the physical similarity. The equivalent (model) materials were selected at random and, as we can see now, not only failed to fulfill the conditions of physical similarity but sharply contradicted them.

Attempts at theoretical generalization in the physics of tectonic deformation were commonly unsuccessful because at that time geologists did not have at their disposal the well-developed concepts of the physics of solids of our day. At that time the science of plasticity, viscosity, and strength of materials was only beginning to develop. Against this background, geology developed a system of physical concepts of its own which was widely accepted in American literature. It consisted of a motley collection of poorly understood ideas randomly snatched from the accumulation of physicomechanical data, commonly distorted and misapplied.

A certain improvement came in the late twenties with the work of Hans Cloos [25, 26], when in elaborating an idea enunciated by E. Reyer as long ago as 1892 [32], he pointed out that in experiments reproducing tectonic deformation with a tremendous diminution of size and duration, it is necessary, in order to preserve physical similitude, to use materials many times less viscous than rocks. This consideration remained, however, purely theoretical.

A theoretical analysis of the question of physical similarity as applied to tectonic investigation was initiated by G. Koenigsberger and O. Morath in 1913 [29] and continued by M. Hubbert in 1937 [28]. By that time a certain amount of material on the physical properties of rocks under high pressures and temperatures had been accumulated, and permitted, at least approximately, by use of the criteria of similarity, the determination of the required properties (first of all, viscosity) of model materials. In the early forties, L. Nettleton and M. Dobrin used these data in their experiment on the formation of salt domes with adequate, for that time, conditions of physical similarity [27, 30, 31]. It is true that the conditions were not consistently fulfilled, but nevertheless this experiment represented a forward step. Very recently other American investigators have repeated the experiment in a somewhat more nearly perfect form [T. D. Parker and A. N. MacDowell, 20] and studied the formation of horst and graben structures by the process of uplift and tension [D. B. Currie, 17].

A much more successful and complete development of tectonophysics on the modern methodological basis has occurred in recent years in the Soviet Union. In 1944 the author of this paper established at the Institute of Theoretical Geophysics, a laboratory of experimental tectonics, the first one in our country. The Laboratory of Tectonophysics is now a part of the section of Geodynamics in the Institute of the Physics of the Earth of the Academy of Sciences of the U. S. S. R. The work of M. V. Gzovskiy in this laboratory played a very important role in the development of tectonic investigation and in the application of correct physical concepts to tectonic deformation. Two other laboratories of experimental tectonics have been organized recently, one at the Department of Dynamic Geology, University of Moscow, by the present author, and the other at the All-Union Petroleum and Geological Research Institute, by D. A. Kazimirov.

The branch of tectonophysics which deals with the physical and mechanical properties of rocks at different pressures and the attempts to determine the dependence of these properties on the composition, structure, and geological history of the rocks also has been developed in the U. S. S. R. The greatest success in this field has been achieved by Professor B. V. Zalesskiy and Yu. A. Rozanov at the Institute of Geology of Ore Deposits, Petrography, Mineralogy, and Geochemistry of the Academy of Sciences of the U. S. S. R. and by Professor M. P. Volarovich at the laboratory of the Institute of Physics of the Earth of the Academy of Sciences of the U. S. S. R.

A detailed tectonophysical field study of the structure of ore deposits aiming at reconstruction and analysis of the history of deformation resulting in these structures is being made by A. V. Pek, V. M. Kreyter, F. I. Vol'fson, I. P. Kushnarev, L. I. Lukin and N. I. Borodayevskiy and many other geologists working in various research and educational institutes of our country. Through their efforts, a historical approach to the study of the structure of mineralized regions has been developed, and in a number of cases, the place of mineralization in the general process of development of tectonic and magmatic phenomena has been determined.

Considering the fund of experience in tectonic investigations accumulated in the U. S. S. R., the following evaluation of the

resent achievements and future prospects of ectonophysics may be made.

The criterion of physical similarity appli-able to tectonic experiments worked out by he Soviet scientists B. L. Shneyerson [24], . N. Lyustikh [19], and more completely by 1. V. Gzovskiy [10, 12], place tectonic experi-nents on a firm theoretical basis and show hat close approximation to physical similarity s quite possible. At present it is still diffi-ult to insure simultaneous similarity of all equired properties -- placticity, elasticity, nd strength -- although theoretically this is ossible. Even now, it is admissible and pos-ible to study the processes, separately on nodels, of elastic and plastic deformation and e formation of dislocations. These are quite ufficient for the solution of a broad group of ectonic problems.

Two physical processes are similar if they an be identically described by the same imensionless equations. This is possible hen definite ratios are preserved between e values of the physical parameters enter-ng into the equations. The ratios are condi-ons or criteria of similarity. It has been hown [12] that for modeling slow plastic eformation and neglecting inertia and elastic rocesses the condition of similarity has the ollowing form:

$$\underline{C\eta} = \underline{C\rho C_1 C_t}$$

here $C\eta$ is the ratio of viscosity of the nodel and the original; $C\rho$ is the ratio of ensities of the model and the object; C_1, e ratio of geometrical dimensions of the nodel and the original; and C_t, the ratio of e duration of the process in the model and n nature.

This equation shows that three of the four hysical parameters may be chosen for the nodel arbitratily, but that our choice will etermine the fourth parameter, which, there-ore, cannot be arbitrary. In practice the reedom of choice of the three parameters is mited by various conditions. For instance, would not be practical to select such model atios of densities, dimensions, and times as o make the viscosity of the model so low nat it is impossible to experiment with it. he size of the model and the duration of e experiment must also be kept within easonable limits. Calculations show that for tudying the slow plastic deformation of a nick (several kilometers) sedimentary depos-, with the model between 5 and 20 cm in nickness and the duration of the experiment ot over twenty-four hours, suitable equiva-ent materials are gun oil, rosin or, in ome cases, very soft moist clay. Clay ontaining 40 to 50 percent moisture is a ery good material also for modeling tectonic dislocations.

The main obstacle to exace experimenta-tion is our inadequate knowledge of the me-chanical properties of rocks at different pressures and temperatures. Best known are their elastic properties.

B. V. Zalesskiy, Yu. A. Rozanov, and B. P. Belikov have determined the dependence of strength and Young's modulus on the com-position, structure, porosity, and geologic history of rocks. M. P. Volarovich, Yu. V. Riznichenko and their associates have deter-mined the effect of confining pressure on Young's Modulus of rocks. The hysteresis in rocks, however, has not yet been determined. Still less well known is the viscosity of rocks. The available figures are obviously very inexact, even in comparison with the deter-minations of viscosity of model materials. It is precisely this circumstance that makes tectonic modeling insufficiently exact. But in spite of this inexactness we have the right to regard tectonic modeling as a very power-ful method of investigation which cannot, of course, at present be used independently but which serves as an important aid to the geologic methods of structural analysis in the field. When correctly used, the experi-mental method always reveals in the repro-duced process some of the important phases which may have escaped the geologist's atten-tion; it reveals relationships among different elements of the process and many of its laws which cannot be discovered in any other way.

The specific and undoubtedly positive trend in Soviet tectonophysics is towards investiga-tion of the fields of stress which were active in the earth's crust in past geologic periods and towards the developement of suitable techniques [13, 14]. An important place among the latter is held by optical methods of inves-tigation of stress in models. A great advance in this field was the development in the Insti-tute of the Physics of the Earth of the Acad-emy of Sciences of the U. S. S. R. (by Gzovskiy and Osokina) of a transparent, optically active plastic which makes it possible to study the distribution of stress in models not only dur-ing elastic deformation, as has been done heretofore, but also of plastic deformation [10]. This greatly broadens the scope of experimentation.

Let us review very briefly some of the latest results of tectonophysical investigations in the Soviet Union.

It was known that the deformation of solids may result in either a fracture or a plastic dislocation. The first follows directly after elastic deformation; the second is the result of development and culmination of plastic deformation. It was supposed that both types

occur among tectonic dislocations. This has been fully confirmed and it is possible now to identify the type of a natural tectonic dislocation and to determine its distinguishing morphological features. The two types of dislocation have been reproduced experimentally.

Elastic dislocation occurs instantaneously as soon as the stress reaches the ultimate strength. In this case only one surface of dislocation is formed (concentrated dislocation). Plastic dislocation develops by means of a gradual concentration of initially uniformly distributed plastic flow in one particular zone. This concentration manifests itself first of all in a more intensive plastic deformation in that zone. Depending on the direction of the acting forces, the character of deformation may vary: when a plastic rod or layer is stretched, necking occurs in the zone of flow concentration; when a plastic body of any shape is subjected to shear by the action of a couple, a fold is formed corresponding to what in geology is called a flexure, etc. Further concentration of flow leads to the appearance, in the zone, of a series of isolated small fractures with very small displacements. In the next stage, some of these numerous fractures stop growing and die out although others, as if availing themselves of the displacements abandoned by the latter, grow in all directions and increase in size. As the process continues, the number of such fractures decreases and at the same time their displacements increase until all movement becomes concentrated on a single surface which is the ultimate dislocation.

Inasmuch as the process of concentration of movement in a plastic body subjected to deformation is irregular, faster in some places than in others, we may find in the body at any moment areas in different stages of this process. These areas are intimately related and are spatial continuations of each other. As a result the continuation of a plastic dislocation (in any direction) is always "dispersed" by splitting into more and more numerous small fractures with ever-decreasing displacements and then by passing into plastic deformation without noticeable dislocation of the body. A plastic fault, for example, passes into a flexure which antedates it and reflects, therefore, an earlier stage of deformation; a plastic thrust passes into an overturned anticline, etc. Faults seldom occur single: there is usually a zone of related faults (a dispersed fault).

These concepts give us a better understanding than was possible before of the conditions of formation of tectonic dislocations and of certain specific features of their structure [1, 5, 7, 11]. There is no doubt that these concepts aid in the solution of some practical problems related to the structure of ore deposits.

It has been discovered from models that each dislocation, whether plastic or brittle, whether shear or rupture, does not form at once but is preceded by the formation of small "embryonic" faults which gradually unite, either slowly or almost instantaneously, into a single large fault surface [15, 23]. This surface bears signs of having formed as a result of addition of many individual dislocations with more or less different histories and slightly different positions in space and is therefore irregular. Studies of natural large tectonic dislocations have shown that they commonly are wavy or festooned [9]. Theoretical investigations and experiments with models, aided by these concepts of the mechanism of dislocations, have led to the conclusion that each individual corrugation of a large fault surface reflects an individual "embryonic" dislocation which became curved before uniting with the adjacent growing and curving "embryonic" dislocations (Fig. 1). It has been shown that the curving of dislocations is caused by a change of stress in a given area as a result of the appearance of the dislocation itself [10, 15]. The waviness of tectonic dislocations plays an important role in the structure or ore deposits, dividing the dislocated zone into a series of open chambers and conduits uniting them.

During a geologic study of a single mountainous area of the earth's crust (Kara-Tau Range) which has undergone a prolonged uplift, numerous fracture systems with varying orientation were discovered [17]. Besides the faults visible in the section (Fig. 2), there are later faults transverse to the strike of the uplift and therefore parallel to the plane of the diagram. They are small normal faults. The dislocations shown in the section are of several types; in addition to the usual normal faults at the crest of the uplift, its limbs bear steep thrusts (reverse faults) and peculiar low thrusts with displacement of the overthrust plate in the direction of the axis of the uplift.

Such a variety of fault systems suggests that at times this area of the crust was subjected to forces acting in different directions.

A special study of this problem by reconstructing the former stress fields from the distribution of faults by the photoelastic method using a transparent elastic model, and finally by reproducing the uplift itself on a plastic model, showed that during the process of upwarping, without application of any other forces besides the force acting from below and responsible for upwarping,

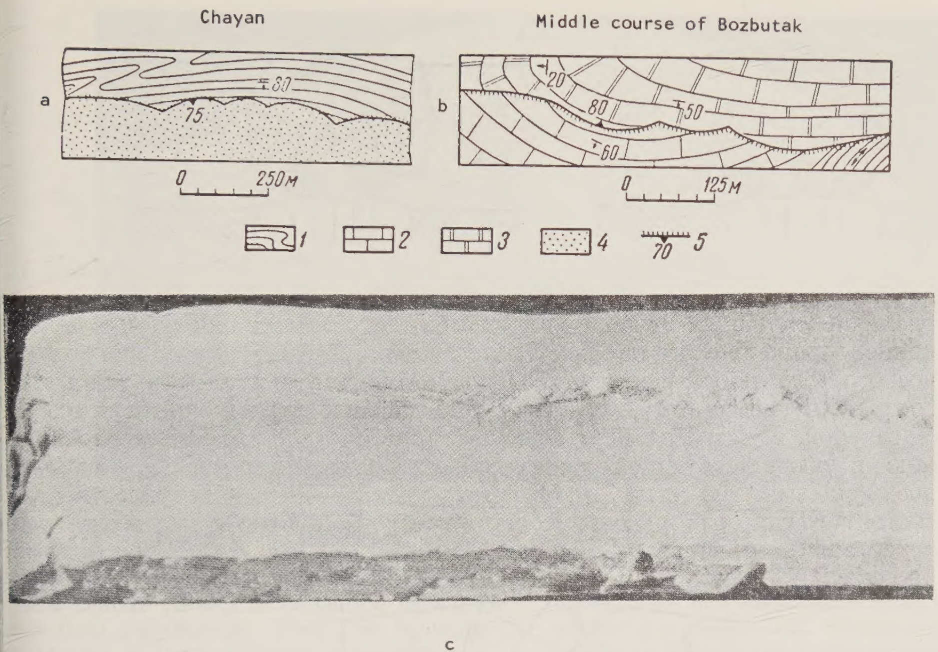


FIGURE 1. Wavy fault surface in nature and in a model

a, b, wavy thrust planes in nature (Kara-Tau Range, M.V. Gzovskiy's observations

1 -- Thin-bedded marly limestones and Lower Tournaisian and Famennian; 2 -- Lower ournaisian and Famennian massive and thick-bedded limestones; 3 -- Lower Tournaisian nd Famennian bedded dolomites; 4 -- Devonian and Siluro-Devonian sandstones and shales; -- traces of large dislocations. Stippling on the lowered block. Triangle and numer, direction and dip of the fault surface.

c, dislocation in a clay model; the zigzag surface of the dislocation can be seen

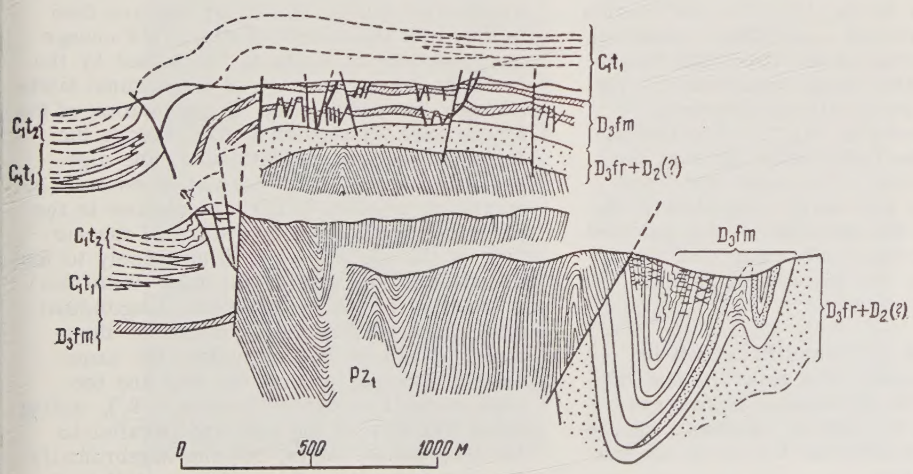


FIGURE 2. Section of uplift in the Kara-Tau Range (after M.V. Gzovskiy)

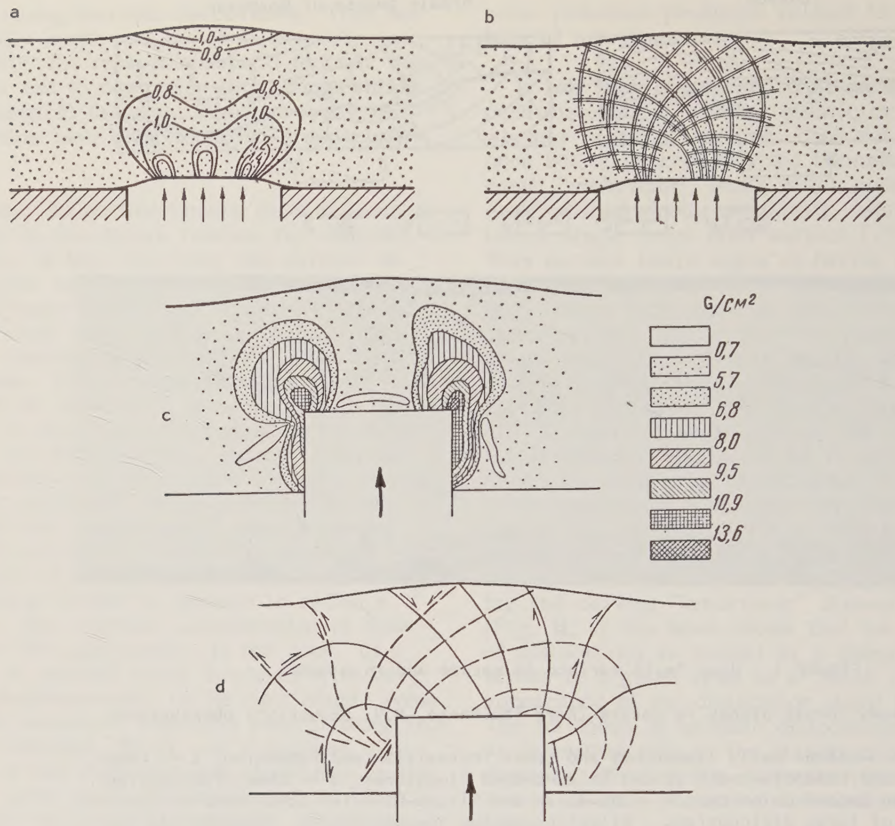


FIGURE 3. Distribution of tangential stresses and trajectories of maximum tangential stresses in models of uplift (after M.V. Gzovskiy)

a -- distribution of tangential stresses in an elastic model; b -- trajectories of tangential stresses in an elastic model; c -- distribution of tangential stresses in a plastic model; d -- trajectories of maximum tangential stresses in a plastic model.

all of the observed dislocations were formed in the upwarped strata [12]. The distribution of tangential stresses under these conditions and the trajectories of the maximum tangential stresses under these conditions and the trajectories of the maximum tangential stresses are shown in Fig. 3. The former indicates that the faults occur first at the base and the crest of the uplift and from there spread up and down, respectively; the latter indicates the direction of the expected displacement in the various parts of the uplift. As we see, the theoretical picture agrees with the observed.

A clay model of the uplift, shown in Figure 4, reproduces the same system of faults. As for the subsequent transverse faults, there is no need to assume a change in the system of external forces to account for their formation.

The models showed that during upwarping caused by the action of a vertical force, longitudinal faults form first and are then followed by transverse faults. This change in orientation of faults is "explained by the fact that the appearance of longitudinal faults changes the possibilities of orientation of the stress axes within the uplift." Before the formation of longitudinal faults, the algebraically largest main normal stresses (i.e., maximum tension, V. B.) are oriented in the plane of the greatest curvature of the surface of the model, i.e., transversely to the direction of the axis of the fold. With this orientation of the stress axes, longitudinal faults form. After their formation, the material can no longer sustain the same tensile stresses across the fold and the main normal stresses (tension V. B.), acting along the axis of the fold and parallel to the longitudinal faults, become algebraically the greatest. It is this new orientation of the stress axes that determines the appearance of transverse faults [12, pp. 542-544].

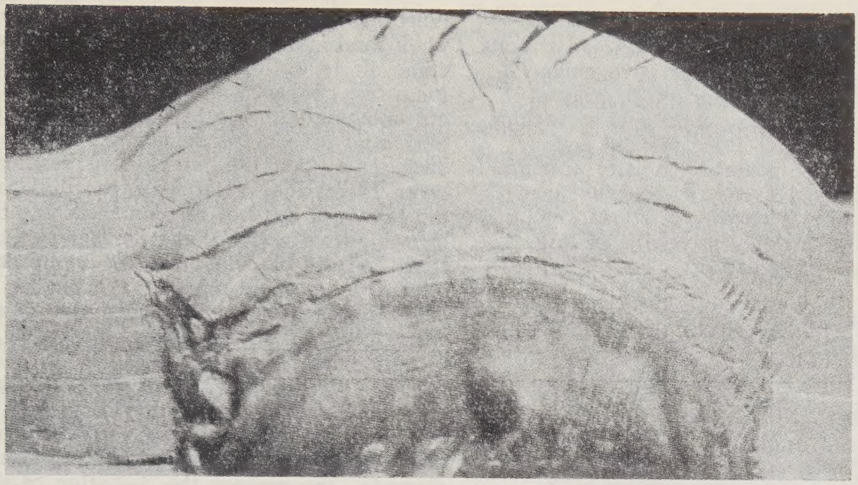


FIGURE 4. Plastic model of uplift (after M.V. Gzovskiy). Moist clay.

The results of this investigation have great ieoretical significance. They give a simple kplanation of the conditions of formation of ie numerous fault systems existing in the irth's crust without the need of using those omplicated and artificial hypotheses requirig repeated changes in the direction of tecmic forces, which are advanced by many eologists.

So-called en echelon fractures are very ommon in nature. Experiments with models have shown that they result from the shearing action of a couple [11]. Models also have helped to explain the s-shaped curvature of the fractures as a result of rotation in the process of continuing deformation of the earlier parts of the fracture together with further growth in the original direction determined by the stress pattern (Fig. 5).

Models have helped to study and explain the phenomenon of boudinage [5, ch. III], which is common in folded regions and

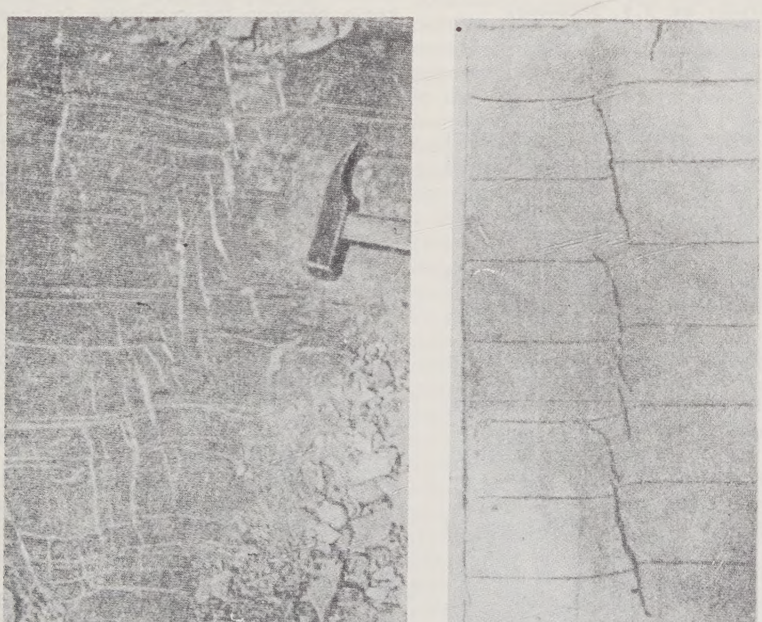


FIGURE 5. Echelon fractures in nature (a) and in a model (b) (after M.V. Gzovskiy)

reveals certain characteristics of deformed strata (the relative plasticity of different rocks under given conditions, the relative magnitude of deformation in different directions, for example, the relative degree of stretching on the limbs of a fold along the dip and strike, the relative general intensity of deformation of individual areas of the crust composed of the same sequence of strata, etc.). A natural and a modeled structure of this type are shown in Figure 6.

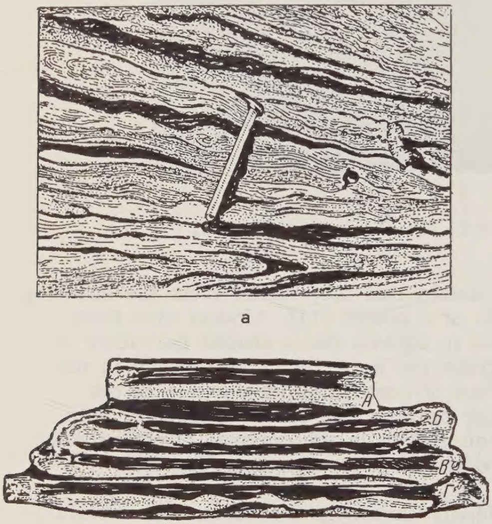


FIGURE 6. Boudinage in nature (a) and in a model (b) (after A.A. Sorskiy and V.V. Belousov)

A, B, C, D -- stages of development.

As a result of these experiments, together with numerous field observations, it has been possible to outline a theory of the formation of tectonic dislocations [1, 5, 12], which, it seems to us, represents an important advance, as compared with the incoherent, poorly reasoned out, and contradictory ideas which have prevailed in this field until now.

One of the central points of this theory is the idea that in all cases dislocations are connected with stresses (maximum tangential and maximum tensile stresses) and not with the magnitude of plastic deformation, as has been supposed (for example, in G. Becker's theory). This greatly increases the scope of the methods of reconstruction of the fields of tectonic stresses formerly existing in the earth's crust from the distribution of faults and the directions of displacement along them [13, 14].

We assume that faults form either as a result of action of maximum tensile stress (rupture) or maximum tangential stress (shear) when the tensile or shearing strength is overcome. It is well known that shear fractures form two conjugate systems which usually do not coincide with the directions of maximum tangential stress but deviate from them in such a way that the principal stress axes become bisectrices relative to the dislocations, the principal axis of tension bisecting the obtuse angles between the conjugate dislocations and the principal axis of compression bisecting the acute angles. The degree of deviation depends on the properties of the material and is determined, apparently, by the effect of normal stresses on its shearing strength. If this effect is absent, a dislocation is formed in the theoretical plane of maximum tangential stress, and the greater the effect of the normal stress on shearing strength, the greater the deviation of the actual shear from the theoretical.

This deviation helps to determine the direction of the principal axes of tension and compression, but it is always desirable to check this by the directions of displacement along the shear fractures.

It is easy to see that this deviation of the shear surfaces from the theoretical position, which has been established by experiment and which follows from the modern theory of the strength of materials, is quite the opposite of the displacement postulated by Becker's erroneous theory. In the latter, the position of shear fractures was connected with the magnitude of plastic deformation and the obtuse angle between the conjugate fractures was supposed to be bisected by the principal compression axis. This question has been fully reviewed by Gzovskiy [13].

Folding is a much more complex process than faulting, and this means that tectonophysical investigations of folding encounter many difficulties.

In order to clarify the questions of mechanism and cause of folding which are still debatable, special investigations of structures in a number of folded regions have been undertaken (by I.V. Kirillova, A.A. Sorskiy, A.V. Dolitskiy, A.M. Shurygin, and V.N. Sholpo in the Caucasus; by M.V. Gzovskiy and V.V. Ez in the Kara-Tau Range; by D.A. Kazimirov in Fergana and by N.A. Syagayev, N.B. Lebedeva and O.M. Filatov in eastern Crimea and northwestern Caucasus). The preliminary results of these investigations show rather definitely that the general crumpling of crustal layers is a result of differential vertical movement of individual crustal blocks. This result is manifested by the movement of material under the force of gravity and the "dynamic squeezing away from the crests" from the relatively raised

lock toward the relatively lowered one. Dur-
ig this process the upper parts of the raised
locks affected by gravity and dynamic squeez-
ig spread fanwise (or mushroomlike) and
ang over the neighboring lowered blocks,
xerting a horizontal pressure upon their
pper parts and causing crumpling of the
trata within a certain belt. These prelimi-
ary conclusions have already been published
l, 6]. They throw an entirely new light on
ie whole problem of the mechanism and
auses of folding.

Until recently, geologists thought that
ılding was due to very general causes on a
anetary scale and did not consider it possi-
e, therefore, to deal with the problem of
ie origin of folding in their studies of a
nall area. It develops now that, in every
dividual case, folding is related to local
uses which can be quite concretely investi-
ated. The basic mechanism of folding is
ot the shortening of the earth's surface but
e elongation of its layers with the preser-
ation of their original area. This elongation
the result of the stretching of the layers
ı the crests of rising blocks and their
eep from the upraised block to the lowered
ie.

In all cases, local causes of folding are
nnected with differential vertical move-
ents of the crustal blocks such as occur
geosynclines, and it is for this reason
at general crumpling is concentrated in
osynclinal zones.

The best material for modeling folding
nnected with gravitational forces is rosin,
hich makes it possible to prepare thinly
yered models and observe the considerable
ructural effects of gravity flow within con-
nient periods of time (hours, a few days).
ie shortcoming of rosin is its stickiness,
hich prevents formation of dislocations, so
at dislocation cannot be observed at the
me time as folding.

Modeling with rosin and blocks which could
moved vertically up or down, arranged
the bottom of the apparatus, thus imitating
e movements of crustal blocks, showed that
changing the sequence of movements of the
ocks and by varying the time between suc-
ssive movements, it is possible to repro-
ce a great variety of fold complexes.
oreover, it developed that if a block is
ised in relation to its neighbors, causing
retching of the layers resting upon it, and
en lowered to its original position, the
retched elongated layers crumple into folds.
hen this operation is performed on several
ocks, an entire folded zone is obtained
hen the blocks are lowered to their original
sition. The crumpling of the layers into
ds occurs, as it should, not as a result
of shortening of the layers but as a result of elongation. The idea that folds may form in this way belongs to D. P. Rezvoy [21]. The whole problem became clear and concrete as a result of experiments with models performed by N. B. Lebedeva at the University of Moscow (Fig. 7).

A study of the details of structure of complete folds suggested certain conclusions concerning the deformations occurring within the layers as they are crumpled into folds. The important role of laminar flow of the more plastic rocks became clear. The great significance of differential pressure normal to the layers also became apparent, for it is this pressure that squeezes the layers and causes the laminar movement of the material [2, 3, 15, 22]. The physical condition of formation of two varieties of folds produced by longitudinal compression -- folds due to bending and those due to squeezing -- also were studied [10].

There existed a conviction that faults (usually reverse faults) were formed in folded regions after the folding, or at least during its later stages, but experiments with models have shown that different relations may exist between the two structures. Thrust faults may form in the earliest stages of folding and even before the appearance of noticeable folds [8]. In such cases folds and faults develop together and the faults, by their position, exert a considerable influence on the position and form of the folds. There is reason to believe that this relationship is common in nature and that some of the peculiarities of folded complexes may be explained only by assuming that the faults appeared before the folds. For example, in the northwestern Caucasus there are numerous longitudinal faults near which large folds disappear (for instance, a syncline may disappear although seemingly it should lie between two anticlines brought in contact by a fault). By using the usual concepts, this relation was interpreted as the effect of a large horizontal component of the displacement along the faults. The faults were pictured as relatively low thrusts concealing some of the elements of the folded structure. However, this interpretation contradicted the general character of the tectonics of the region where the possibility of large horizontal displacements was excluded. If it is admitted that faulting antedated folding, and that the latter formed independently within each area separated by faults without any connection with folding in adjacent areas, then there is no need to seek geometrical perfection in the structure and look for the location of "lost" parts of the folds. These parts never existed. At the same time, the faults may be interpreted as normal faults, which accords better with the general

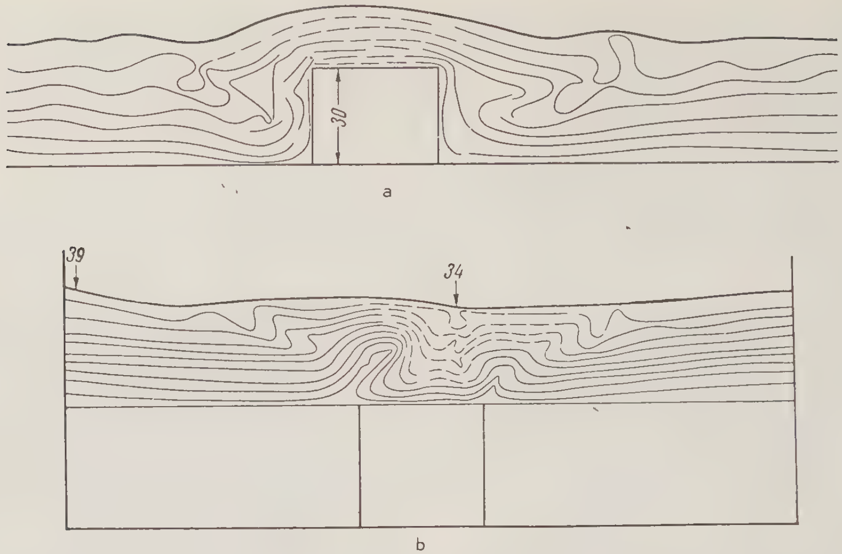


FIGURE 7. Experimental reproduction of regional folding by means of vertical movements of blocks and elongation of layers. The material is rosin with partings of clay. After N.B. Lebedeva.

a -- deformation of rosin layers by the movement of one block; duration of experiment, 18 hours; b -- deformation of rosin layers after raising the block 30 mm and then lowering it to the original position after 24 hours. The structure becomes stable after 26 hours.

All diagrams show the surface of the model and the position of clay partings separating rosin layers. Where clay layers are shown by an interrupted line, the layers of rosin were stretched and elongated.

structure of the region.

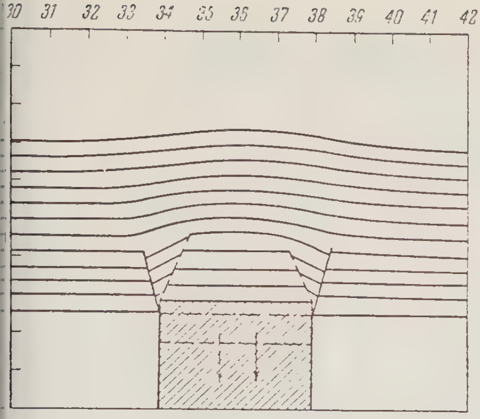
The results of modeling of the so-called block folds, especially of reflected block folds, i.e., folds formed directly by pressure induced by the rising of an isolated crustal block, are particularly interesting. It was found, for example, that if the deep layers reflecting the shape of the crustal block (i.e., of the block in the model) are crumpled into box folds, then upwards in the series of layers they become gradually gentler and broader and their resemblance to the upper surface of the block decreases. A detailed study of this phenomenon on models was made by A.M. Sycheva-Mikhailova. If the block of the model has an asymmetrical form (one-sided uplift) only the lower layers reflect this asymmetry and form a steep flexure over the edge of the block. The asymmetry disappears very rapidly and higher in the section the layers are bent into a low symmetrical swell. In the same manner, negative forms (downwarps) broaden and die out upwards in the section. In thick series of layers a generalization occurs, as it were, and two deep-seated uplifts unite on the surface into one if the thickness is great enough. The layers will dip steeply at the surface of a thick plastic mass only along the seam separating two blocks moving in opposite directions, one rising, one subsiding.

The faults which accompany the movements of the blocks are differently distributed on the rising and subsiding blocks. From the edges of a rising block, reverse faults converge sharply, producing over the block a truncated pyramid pointing upward. Within the pyramid, the layers are horizontal and do not participate in the deformation occurring on the limbs of the uplift.

Over the edges of a subsiding block the reverse faults diverge upwards. If the same block is first raised and then lowered, both systems of reverse faults form in succession and when the block is returned to its original position, two faults spring from each of its edges.

These results are shown in Figure 8.

Pressure folds produced by the squeezing out of plastic material from beneath structural depressions (due to the unequal distribution of weight of the overlying layers) towards anticlines have also been reproduced in models to aid in field investigations.



—·— Original position of the block.
——— Position of the lowered block.

IGURE 8. Model of block folds. Faults after
uccessive lowering and raising of the block.
A.M. Sycheva-Mikhaylova's experiment.

First the process of formation of salt ugs was modeled by "floating" a lighter scous material which represented salt thin a heavier viscous material which itated the enclosing rocks [18]. The tech- que of the experiment is simple: crude l of low specific gravity was poured on e bottom of a vessel and over it a layer heavier transparent sugar syrup. Through e latter it was easy to observe the process the rising of the oil in the form of black lumns and to establish a relationship be- een the rate of growth of the plugs and e form, on the one hand, and the differ- ce in density of the two materials, on the her. If the thickness of the lower viscous quid ("salt") was different in different aces, the diapirs formed first of all, or clusively, in the zones where it was thick- t, which agrees with the localization of lt plugs in the depressions in the relief of e rocks underlying the salt, i.e., in those aces where the salt beds are thicker. The te of growth of diapirs is increased even ore by the increase in the thickness of the erlying heavy liquid. That the growth of apirs is favored by the great thickness of erlying sediments has been observed in ture. The growth of a large group of multaneously or successively forming dia- rs also has been observed in models. This as done with an apparatus of large area ing modeling material of sufficiently high scosity to slow down the process. The odeling of the growth of diapirs in groups d never been done before, for it was not tempted in the experiments of American vestigators.

The process of formation of domes with clay plugs also was reproduced with models. The clay diapirs of the Kerch-Taman' region were studied especially for this purpose and served as prototypes. The experiment was based on the hypothesis that plastic clays form domes by being squeezed out from under downwarps by the weight of overlying rocks. A typical "piercement" dome formed in the models when the material imitating plastic clay had free access to the surface through an opening (crack) in the overlying material. It may be supposed that the final shaping of the "clay diapirs" in nature also occurs after the plastic clay mass (in the case of the Kerch-Taman' region, the Maykop formation) comes to the surface through an opening eroded in the crest of the uplift which originally had, perhaps, the form of a block fold. It developed that the structure of the plug depends on the width of the opening through which its material was protruded, and on the distribution of load in the downwarp from which the clay was squeezed out. In some cases the plug has a complex folded internal structure; in others, the form of a single anticline with strongly stretched layers (the stretching of the layers is indicated by the breaking of the very thin clay partings between the more plastic layers of petroleum or rosin).

It is very interesting that under the conditions of the assumed mechanism the upper layers of the material imitating clay beds come into motion first. Because of this, the peripheral parts (limbs) are thrust against the inner part in the squeezed-out plugs. This fully reproduces the structural features which have been revealed by boring of certain comb-like anticlines of the Tersk-Sunzha region of northern Dagestan. It may be supposed that these anticlines also were formed by the flow of plastic clays from underneath downwarps overloaded with sediments. The last phenomenon may be the explanation of the superimposed synclines so widely developed in the crests of anticlines on the Kerch Peninsula. The more intensive squeezing out of the Maykop clays upwards on the limbs of the folds must lead to the formation of secondary synclines along the axial zones of the folds and to the preservation here of the layers overlying the clay. Moreover, being heavier than the water- and gas-saturated Maykop clays, these layers must have sunk into the clays and the sunken masses gradually acquired rounded, basin-like shape, mechanically the most advantageous (Fig. 9). These results of study of the mechanism of formation of clay diapirs in the field and on models were obtained by N. B. Lebedeva at the University of Moscow.

The results of the tectonophysical experiments cited above convince us that this approach is a very important addition to

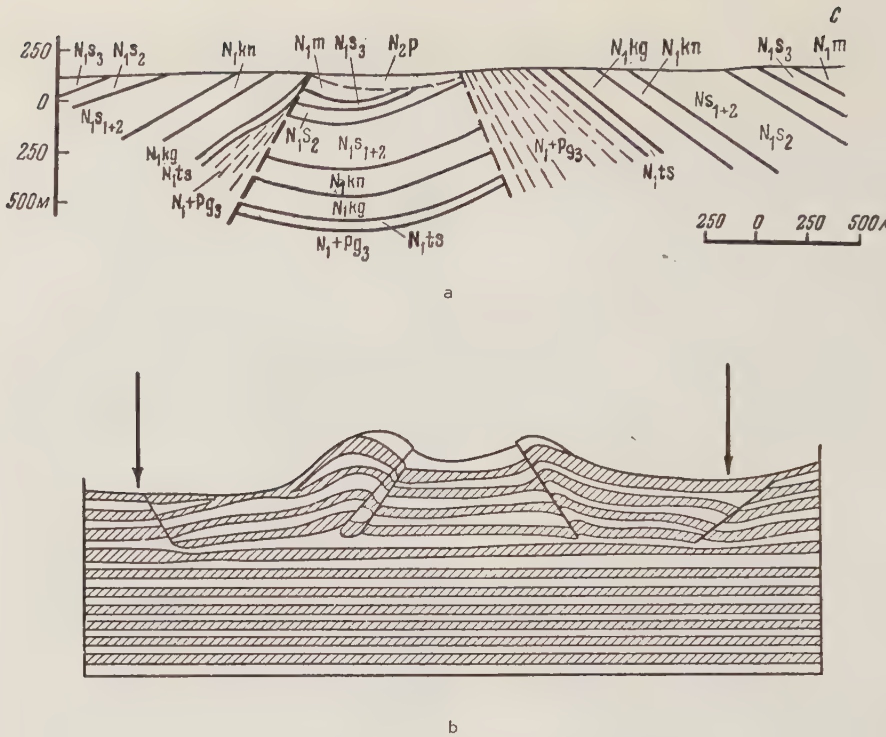


FIGURE 9. Diapir folds in nature and in model

a -- Section of the Andreyev Diapir fold (Kerch' Peninsula); b -- model of a diapir fold. Material -- gun grease (after N.B. Lebedeva)

strictly geologic methods. Many of these experiments reveal and explain those phases of tectonic phenomena which are not accessible to the purely geologic methods.

The following conclusion can be made on the basis of accumulated experience with tectonophysical investigations.

Such investigations should not be incidental or made sporadically. In order to achieve success in this field, specialization and long systematic effort are necessary. This statement is pertinent, for in the United States, for instance, a country which is considered the center of modern development of tectonophysics, investigations in this field are conducted on randomly selected problems by different investigators and without a guiding plan. The same attitude towards tectonophysics existed until recently in our country. An organization of the broadest possible program of detailed field studies is essential, for it is absolutely impossible to attack the problem of folding, for example, on the basis of random brief observations made now in one region, now in another. There must be an appropriate choice of the object of study and a consistent effort dictated by the problem itself and not by other considerations.

We are a great deal behind as far as experimental studies of the mechanical properties of rocks under high pressure and temperature are concerned. Only when these properties become known will it be possible to increase the exactness of tectonic modeling. At present, we can reproduce tectonic processes only qualitatively, in a generalized way, since many details must be neglected because they are beyond the scope of experiment; but when properties of rocks become better known, it will be possible to reproduce specific structural forms observed in nature, and this, of course, will be extremely valuable.

It is the plastic properties and strength of rocks that must be known within a broad range of pressure and temperature, and also the effect of the rate of deformation on these two properties.

Almost all work in tectonophysics is conducted at present by geologists who must at the same time be physicists. It is very desirable that physicists also become interested in this field of study where, with obvious usefulness to the physics of the solid state, they could study large plastic deformations in vast volumes of heterogeneous

material developing over long geologic periods of time. The combined work of geologists and physicists would yield, undoubtedly, very valuable results.

Tectonophysics is one of the borderline fields of knowledge. It is well known that it is from such borderline fields that modern science receives the impetus for further advance. It is here that refined methods of investigation are developed and new points of view on well-known processes are provided; it is here that ignorance and scientific prejudices are destroyed. Tectonophysics possesses all of these advantages and possibilities. It represents that "point of growth" which in the near future will be able to exert the deepest influence on the development of basic concepts in the realm of the most important theoretical and practical problems of geology and associated sciences.

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CERTAIN PROBLEMS OF THEORETICAL VOLCANOLOGY

by

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The author's observations of the screening of transverse seismic waves lead him to the onclusion that the reservoir of molten magma beneath the Klyuchevskiy volcano lies at a epth of the order of 60 km, i. e., at the boundary between the earth's crust and mantle. n the basis of observations on the form of the "seismic shadow" the author presents his eas about the size and shape of the magmatic chamber.

* * * * *

I

One of the fundamental problems of odern volcanology is the question of the epth of volcanic magmatic chambers. A orrect solution of this problem would prode the solution of many related problems ot only in volcanology but in dynamic eology, in the theory of ore deposition, nd in the tectonics and physics of the irth.

For a long time there has existed in olcanologic literature the concept of shalw magma chambers. This traditional conept is based on various more or less arbiary postulates and not on geophysical oservations. The author attempted to estiate the depth of the magma chamber of e Klyuchevskiy volcano by using the oserved phenomenon of screening of transerse seismic waves [1].

During seismological observations begun 1948 at the Kamchatka Volcanologic Staon of the Academy of Sciences U. S. S. R. t Klyuchi), seismographs registered many undreds of local and distant earthquakes. ne local earthquakes do not differ from milar earthquakes in other regions; the rrival of $\bar{P}$, P*, and P waves and of $\bar{S}$, S*, nd S waves from shallow foci proves that e structure of the earth's crust in the egion of Kamchatka is in no way different om that of other continental areas and at both granitic and basaltic layers of onsiderable thickness are present.

The records of the majority of distant irthquakes show the arrival of P and S aves quite clearly, indicating that there is continuous layer of liquid magma beneath e crust which if sufficiently thick would screen off all transverse waves. Only certain Japanese earthquakes provide an exception to this. Earthquakes occurring on Hokkaido and northern Honshu are recorded almost fully, but records of earthquakes with epicenters in southern Japan, of which more than twenty have been registered at the Klyuchi station show longitudinal waves only (P, PP, and others), whereas the direct transverse S waves are absent. Earthquakes occurring on the same azimuth as Japan but farther away (Philippine Islands, Celebes) show clear arrival of S waves.

A possible cause of nonarrival of S waves might be the unfavorable position of the station with respect to the forces at the focus. However, as the latest investigations show [3, 8, 9], most earthquakes in the region of East Asia island arcs are due to thrusts in the direction almost normal to the continental margin, i. e., in the direction most favorable for recording at Klyuchi. At the Petropavlovsk station, which lies approximately on the same azimuth from Japan as Klyuchi, the S waves are sharply recorded. It is quite evident that the suggested cause is inadequate.

The possibility of a sharp decrease in wave energy at a given epicentral distance has not been analyzed in detail, because earthquakes occurring in other regions at the same epicentral distance as Japan (for example, Alaska) give a sharp S wave record. The Magadan station, located at the same distance from Japanese epicenters as Klyuchi, records the arrival of S waves clearly. It is evident that the epicentral distance is not the cause of nonarrival of S waves at the Klyuchi station. Fig. 1 illustrates the above statements.

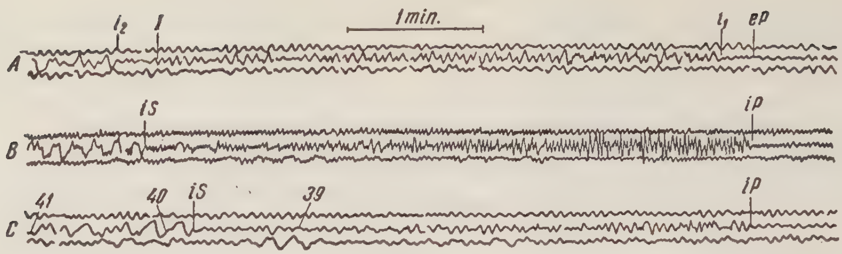

FIGURE 1. Seismograms of the Japanese earthquake of November 27, 1953, 11h 30m 08s = 2 sec. NE component.

A. Klyuchi Station, $\Delta = 25^{\circ}$; 6eP, 11h 35m 36s; i_1, 11h 35m 50s. I. 11h 39m 59s expected arrival of S waves computed from the time-distance graph; i_2, 11h 40m 17s. B, Magadan Station, $\Delta = 26^{\circ}$; 1 iP, 11h 35m 39s; iS, 11h 40m 06s. C. Petropavlovsk Station, $\Delta = 25.2^{\circ}$; iP, 11h 35m 04s; iS, 11h 39m 10s.

It is clear that the absence of transverse waves from the record is due to a purely local cause, to the existence of a "screen" in the vicinity of the station. The epicenters of earthquakes for which S waves are not recorded lie at distances ranging from 24° to 50° on the azimuth of 214° to 230° from Klyuchi. This direction passes through the Klyuchevskiy group of volcanoes. The most natural and likely cause of seismic shadow for the direct transverse waves is their screening by the liquid magma of the magma chamber.

Knowing the distance to the summit of the volcano, i.e., to the volcanic vent connecting the magma chamber with the surface, and having determined the angle of emergence of the screened waves, it is easy to calculate the depth of the "screen", i.e., of the magma chamber. The angles of emergence of P waves for different epicentral distances are given in many texts, and to simplify calculations, it is usually assumed that the longitudinal and transverse waves travel the same route. This, however, is not so, for their velocities are different and their angles of emergence for a given epicentral distance must also be different. Therefore, the existing tables and graphs of the angles of emergence of P waves cannot be used and the angle of emergence of S waves must be determined.

This determination may be made by differentiating the time-distance curve:

$$\cos e = \frac{dT}{d\theta} V_s, \qquad [1]$$

where e is the angle of emergence, T, travel time of the wave over the distance from the epicenter θ, and V_s, the velocity of the transverse wave near the earth's surface.

The results of computation of the angle of emergence of S waves are presented in the graph of Fig. 2, together with the curve of the angle of emergence of P waves. It will be seen that the difference in the angles of emergence of P and S waves varies significantly with the epicentral distance.

It should be added that numerical differentiation of experimental curves, such as time-distance curves, does not give absolutely exact results and that the simpler method of using a finite ratio of increment instead of the derivative is sufficiently accurate for our purpose. Substituting the numerical value $V_s = 3.5$ km/sec in equation [1], we obtain

$$\cos e = 3.5 \frac{\delta T}{d\theta} \qquad [2]$$

or, changing kilometers to degrees:

$$\cos e = 0.031 \frac{\delta T}{\delta \theta^{\circ}} \qquad [3]$$

The transverse waves at the Klyuchi station are screened when the epicentral distance ranges from 24° to 50°. According to Fig. 2, the angles of emergence of S waves corresponding to these distances are from 57.5° to 64.5°.

The depth of the "screen" H, with the horizontal distance to it being l, is:

$$H = l \tan e, \qquad [4]$$

and, since the distance from the station to

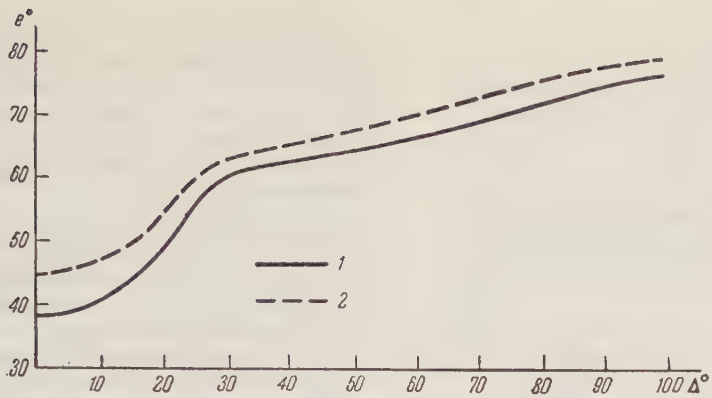

FIGURE 2. The dependence of the angle of emergence of seismic waves on the epicentral distance.

1 -- Angles of emergence of transverse waves. 2 -- Angles of emergence of longitudinal waves. 3 -- e-Values of the angles of emergence, Δ-epicentral distance.

e Klyuchevskiy volcano is 33 km, the pth of the magmatic chamber must be tween 50 and 70 km; i.e., it lies almost the boundary of the earth's crust and antle. Let us recall that recent work on e eastern margin of Asia indicates that e crust beneath it is as much as 60 km ick.

Examination of the boundaries of the eismic shadow" gives some idea of the ze and shape of the magmatic chamber. ie "seismic shadow" has the form of an liptical cone making an angle of 60^{o} with e earth's surface. The angle between the nerators in a vertical section (α) is 7^{o} d in a horizontal projection, 15^{o} to 17^{o}.

From this it may be deduced that the amber has the form of a convex lens, perhaps of a triaxial ellipsoid elongated the east-west direction and flattened in e north-south direction. The length of e chamber and its thickness is estimated 25 to 35 km and its volume as 10 to 20 ousand km^3.

The actual form of the chamber may, course, be much more complex and clude various projections and apophyses.

II

Many records of Japanese earthquakes ow the arrival of the secondary wave 2, Fig. 1) 8 to 14 seconds after the comted expected arrival time of the *S* wave. is wave is considered as a converted ve refracted by the magmatic chamber the *SKS* type) and is designated *SRS*. idently the sharp arrivals recorded 14 19 seconds after the arrival of the diffracted *P* wave (*PRP*) are of the same nature (i_1, Fig. 1).

The difference in arrival time of direct and converted waves makes it possible to determine the velocity of the converted wave in the magmatic chamber. It is easy to see that:

$$V_R = \frac{VH}{H + V \Delta t} \qquad [5]$$

where V_R is the velocity of the converted wave in the magmatic chamber; V, the velocity of the direct wave near the boundary of the magmatic chamber; H, the length of travel of the converted wave in the magmatic chamber; and t, the difference in times of arrival of the direct and converted waves. H = 30 to 35 km, V_P = 7.8 km/sec, V_S = 4.6 km/sec, Δt is 12 to 14 sec for *S* and 18 to 19 sec for *P*.

From these data the value of V_R is between 1.6 and 1.8 km/sec, or near the velocity of longitudinal waves in water or in unconsolidated sediments (under normal pressure).

Considering that the shear modulus μ of the material in the magmatic chamber is zero or near it, we have:

$$V_R = \sqrt{\frac{K}{\rho}}, \qquad [9]$$

where K is the modulus of compression of the material in the magmatic chamber and ρ is its density (3.4 g/cm^3), i.e.

$$\underline{K} = \rho V^2 \quad [7]$$

whence $K = 1 \cdot 10^{11}$ dynes/cm^2 or $1 \cdot 10^5$ bars.

The reciprocal magnitude, compressibility (β) is $10 \cdot 10^{-6}$ bars^{-1}. This value for compressibility is the average of its values for a typical solid and a typical liquid (for the earth's crust $\beta = 0.8 \cdot 10^{-6}$ bars^{-1}; for water, $\beta = 48.9 \cdot 10^{-6}$ bars^{-1}).

Thus the study of seismograms gives a certain idea of the elastic constants of the material in the magmatic chamber.

III

These deductions concerning the depth and form of the magmatic chamber and the elastic constants of its material must be regarded as very tentative. However, the considerations presented above are sufficient to serve as a basis for a brief review of some general questions of volcanology.

1. The location of magmatic chambers at the boundary of crust and mantle or, what is more probable, in the upper parts of the mantle is by no means accidental. It is précisely at these depths (80 km) that B. Gutenberg [6] has recently discovered a decrease in the velocity of seismic waves and therefore a decrease in elastic constants. The most probable cause of this phenomenon is the change from the crystalline to the amorphous state. On the other hand, E.A. Lyubimova [4] has shown theoretically that in the upper part of the mantle the geothermal curves and the melting curves of materials almost coincide. Therefore, conditions must exist in the upper part of the mantle which are particularly favorable to fusion of rocks with change in thermodynamic conditions. Large tectonic dislocations accompanied by local decrease in pressure of increase in temperature must inevitably lead to the formation of magmatic reservoirs.

From this follows the natural connection between modern volcanism and the zone of Alpine orogeny.

2. In some cases the molten magma is pressed upwards but does not reach the earth's surface and freezes in the form of basic dikes and sills. In the localities where the earth's crust is deeply fractured (down to 60 to 80 km), the liquid magma reaches the surface in one of the forms of effusive volcanism. Volcanoes are always related to deep fractures in the crust and it is immaterial whether these fractures are in a zone of recent orogeny or whether they pass through an already consolidated platform or an ancient mountain region. In every case, deep fracturing (60 to 80 km) serves to localize volcanic activity, as shown by the presence of modern volcanoes in Manchuria, Tibet, and eastern Africa and of very young volcanoes in northeastern Asia.

3. The localization of magmatic chambers at a considerable depth where earth materials have no sharp chemical differences explains very simply the amazing similarity of lavas within extensive regions. Small magmatic chambers could not account for the uniformity of lavas over many thousands of kilometers of the geosynclinal zone on the periphery of the Pacific Ocean. At the same time, chemical variation in lavas in the direction normal to the continental margins testifies to considerable difference in chemical composition of the upper part of the mantle at the lower part of the crust, depending on oceanic, geosynclinal, or platform character of the crust or, in the final analysis, on the basic geologic process of formation and degree of development of geosynclines. Therefore, the lavas of the continental volcano Uyun-Kholdonga, in Manchuria, are close to the lavas of the continental volcanoes of remote eastern Africa and are quite different from the nearby lavas of the geosynclinal volcanoes of Japan.

4. An important role in theoretical volcanology is played by the hypothesis of peripheral magmatic chambers. This hypothesis, advanced by A. Kircher [9] in the seventeenth century, was revived in the beginning of our century by Stubel [10]. Stubel's ideas on the development of volcanoes have been abandoned, but the idea of peripheral magma chambers continues to dominate many minds.

We shall not discuss this idea in detail. Let us say only that during earthquakes in the south Kamchatka region and in the northern Kurile Islands, the waves which pass through the Klyuchevskiy volcanic group at a depth of about 30 km do not exhibit any anomalies indicating that at this depth there cannot be any large volumes of molten magma.

Local earthquakes in the region of the Klyuchevskiy volcano, with short periods and shallow foci (3 to 10 km), also show no anomalies. The $\underline{S}$ waves from local earthquakes have periods of 0.2 -- 0.4 sec, and would be readily screened by a liquid layer 500 to 1,000 m thick. Thus seismological data show the impossibility of the existence of peripheral chambers at shallow depths. This conclusion must be

;enerally true. Exceptions to it are ex-remely rare and may be only the result of netasomatic replacement of carbonate rocks, .s appears to be the case at Vesuvius.

5. In eruptions occurring at intervals of ens or a few hundreds of years, a sharp ifferentiation of lavas is usually observed; he eruptions begin with acid lavas and end /ith basic (Krakatoa, Hekla). The total olume of products of one eruptive cycle eldom exceeds a few cubic kilometers. 'he volume of a volcanic vent with the ength of 60 km is at least 60 to 80 km^3. : is quite evident that rapid and sharp hemical differentiation of magma between aroxysms occurs in the volcanic vent.

The slower and less sharp differentiation 1 the magmatic chamber trends towards 1crease in alkalinity and takes place over 1ousands and tens of thousands of years. .n occurrence of both types of differentia-:on was brilliantly illustrated by Zavarit-kiy by the lavas of Vesuvius [2].

6. Nonsynchronous eruptions of volcanoes f the same group result from processes ithin the volcanic vent. Such eruptions ither occur randomly or each volcano may xhibit its own rhythm which does not epend on the rhythms of its neighbors. uch eruptions are accompanied by earth-uakes with shallow foci lying, usually, at epths of 3 to 5 km.

Synchronous eruptions of many volcanoes f the same or several volcanic regions re due to profound regional, and sometimes erhaps, planetary causes affecting a num-er of magmatic chambers simultaneously Andes, 1932; Kamchatka and the Kuriles, 945-1946). These causes are not yet nderstood but they are probably related to ie activation of tectonic movements. This : suggested by the fact that earthquakes of itermediate focal depth of 60 to 100 km ommonly precede such synchronous erup-ions.

The appearance of new volcanoes (Pari-utin) or the awakening of dormant ones lezymyannaya Sopka in Kamchatka) is also elated to activation of the processes in ie magmatic chamber and is also accom-anied by intermediate-focus earthquakes.

8. The position of a magmatic chamber : a considerable depth and the absence of eripheral chambers narrow the possible olutions of certain problems of theoretical olcanology. For example, it becomes lear from this point of view that the forma-on of calderas cannot be due to the col-ipse of the roof of the magmatic chamber. must be admitted that the old hypothesis of Escher [5] is correct, i.e., the calderas are formed as a result of collapse of the walls of the vent emptied by the preceding eruptions.

* * *

These examples suggest the possibility of creating a general volcanologic theory, based on the knowledge that magmatic chambers are located at great depth, which would generalize a number of basic and particular manifestations of volcanic activity.

It is highly desirable to check the author's conclusions concerning the depth of magmatic chambers by deep seismic sounding.

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EPIGENETIC FEATURES OF SANDSTONES OF THE MOGILEV FORMATION OF THE SOUTHWESTERN PART OF THE RUSSIAN PLATFORM AND CERTAIN PROBLEMS ASSOCIATED WITH THEM

by

A. V. Kopeliovich

The study of drill cores from the southwestern part of the Russian platform establishes ofound textural alteration of the arenaceous Mogilev sediments which are genetically lated to the formation of a definite assemblage of secondary minerals. This alteration aracterizes a late epigenetic stage and does not differ from the changes occurring during e early stages of metamorphism.

* * * * *

The oldest rocks of the sedimentary antle of the southwestern part of the issian platform have been studied by nu- erous investigators for more than sixty ars. The problems of their stratigraphy, e, and mode of formation have been peatedly discussed in published works. nsiderably less attention has been paid their composition, mineralogy, and ructural features. Our investigations have own that the rocks composing the lower visions of the section in the southwestern rt of the Russian platform are very dis- ictive. They show signs of intensive eration which has substantially modified th their mineral composition and their ructure, especially in the deeply sub- erged parts of the platform. In the auth- 's opinion, an investigation of the char- ter of these alterations will establish rtain features of the epigenetic processes general and throw light on the problem the change of sedimentary rocks into etamorphic.

In this paper we shall discuss the fea- es of the secondary processes in the ndstones of the Mogilev formation only, , in it because of the relatively simple neralogic composition, the results of se processes are most clearly expressed.

The Mogilev formation lies at the base unfossiliferous terrigenous deposits of e-Gothlandian age of the Dnestr region. rests on the crystalline basement com- sed of different gneisses, granites, and, re rarely, of basic rocks, and is over- n by argillites, siltstones, and sand- nes of the Ushitsa formation, which, in its turn, is transgressively overlain by the Molodovo sandstones containing a Caradocian fauna [16]. The age of the Mogilev formation has not been established, some geologists referring it and the entire pre-Gothlandian terrigenous series of podolia to the Ordovician [10, 16], others, to the Cambrian [1, 6, 7], and still others, to the Sinian and Riphaean [5, 9, 11].

In the investigated sections, the Mogilev formation lies at considerable depth, in the Mirnoye borehole (Odessa region) in the interval between 1611 and 1544 m, and in the Kaushany borehole, in the interval between 1400 and 1385.5 m.

The formation consists of conglomerates and sandstones with subordinate beds of micaceous sandstones. In the Odessa section, in the middle part of the section, among the sandstones there is a group (16 m) of micaceous siltstones and argillites.

I. THE ORIGINAL COMPOSITION AND ASPECT OF THE ARENACEOUS ROCKS OF THE MOGILEV FORMATION

The pebble conglomerates and sandstones are yellowish-gray, pink and reddish, sometimes with a greenish cast. The differences in color result from variation in the proportion of dark-gray quartz, yellowish and white plagioclase, rose and red microcline, and greenish micas. In the Mirnoye borehole, the color of sandstones in some beds is brown, because of the presence of iron

hydroxides in the cement.

The sandstones are predominantly very coarse and coarse grained, commonly inequigranular, sometimes fine grained with an admixture of granules of quartz and feldspar and, in few places, of pebbles of granitic rocks. The structure of the conglomerates and sandstones is massive, coarsely laminated, rarely finely laminated. The lamination is due to the difference in granulometric composition of individual laminae, to the orientation of elongated detrital grains and flakes of mica, and in places to differences in color. The lamination is usually horizontal. In the sandstones of the Mirnoye borehole, V.N. Kortsenshteyn [1] noted curved, lenticular cross-lamination and wave ripple marks.

The rocks are hard and unusually dense. Their porosity ranges from 1.9 to 5 percent, averages 2 percent and approaches the porosity of igneous and metamorphic rocks. The sandstones and conglomerates are often pierced by vertical threadlike veinlets of quartz and ankerite.

In composition these rocks are typical arkoses. They contain grains of quartz, microcline and plagioclase, and flakes of mica. Occasionally there are lithic fragments of quartzites and, less frequently, of granites. The accessory minerals are zircon, tourmaline, magnetite, and in places garnet and epidote.

Quartz is predominant in the rocks and usually contains extremely small liquid and gas inclusions arranged in chains. The larger quartz grains have wavy extinction, passing in places into mosaic extinction. Some quartz grains are recrystallized into aggregates with an intricate pattern.

The feldspars are present in considerably smaller amounts (20 to 25 percent). Predominant among them are microcline and microcline perthite, usually with the grid pattern and either water-clear or turbid with clay. In some grains there are radiated aggregates of large kaolinite flakes. The plagioclase, represented by albite-oligoclase and sodic oligoclase, is polysynthetically twinned and is usually weakly, but may be rather strongly, sericitized.

The quartzite fragments, with characteristic serrated boundaries, is usually present in small amounts (3 to 8 percent). Only in the conglomerates does their content rise to as much as 12 to 15 percent in some layers.

The micas are represented exclusively by biotite in small and large altered flakes with residual pleochroism in brown (Kaushany) and green (Mirnoye) tones. The content of biotite varies from 1 to 10 percent.

The detrital grains, where their original shape has been preserved, are equidimensional, somewhat elongated, and subrounded to subangular. The coarse sand grains (0.5 mm) and granules are usually well rounded.

The detrital material is sometimes well sorted and sometimes poorly sorted. The better sorted medium- to fine-grained sandstones are usually better sorted and occur, as a rule, in the upper parts of the formation, while the coarse-grained varieties and granule conglomerates containing a considerable proportion of fine grains as well as pebbles compose the lower part of the section.

II. ALTERATION OF THE ROCKS

The sandstones and conglomerates have undergone intensive alteration. In a number of cases it is so great that the original detrital form of the grains is partly or completely lost and the texture of the rocks in some areas is blastopsammitic, in others quartzitic or crystalloblastic and resembles more the texture of metamorphic rather than sedimentary rocks. The profound textural modifications are accompanied by a noticeable grouping of material with replacement of some minerals by others and the appearance of parageneses different from those of the original rocks.

A general tendency is detected in these varied and complex processes which is manifested in the development of textures with the closest packing of grains, high density, and minimum porosity. These textural changes are genetically related to the formation of new mineral associations which are stable in the physicochemical environment of metamorphism.

The alteration of the rocks resulted from two opposed processes, solution of detrital grains and crystallization of new minerals from these solutions. The intimate intermingling of these processes led to the phenomena of replacement and development of crystalloblastic textures.

1. Structures Due to Solution

Among the structures due to solution in the sandstones of the Mogilev formation, microstylolite and structures of adaptation

(conformation) and penetration (incorporation) are widely developed.

The Microstylolite Structures are best developed in the rocks of the Mirnoye section. In the sandstones of the Kaushany borehole they are found less commonly and are less typical. These structures are almost invisible to the naked eye and are revealed only in thin sections under the microscope. The reason for this is not only the small size of the microstylolite seams but also their light color, invisible against the light background of the rock.

In the sandstones and conglomerates of the Mogilev formation there are the following types of microstylolites: 1) microstylolite surfaces of considerable extent, 2) microstylolitic grain contacts and, 3) microstylolites within the grains.

Microstylolitic surfaces are rather seldom detected under the microscope, apparently because the rock splits easily along them and the probability of such a surface being preserved in a thin section is small.

In cross section, microstylolite surfaces appear as thin (from thousandths to 0.1 mm) sinuous or serrated seams along which one part of the rock penetrates the other in the form of teeth, knobs, or irregular, sometimes intricate shapes ranging in height from 0.05 to 0.55 mm and occasionally up to 1.5 mm. The length of the seams exceeds the length of a thin section (Fig. 1). The microstylolite columns vary in form. Sometimes they are rather regular columnar or prismatic structures, developed in quartz or microcline and penetrating into a neighboring grain; sometimes they are of acidular, stalactite-like, wedgeshape, or knoblike shape, or form protuberances with oval outlines. Columns consisting of aggregates of small quartz and feldspar grains squeezed into another grain or into another aggregate are commonly observed, and in such cases there are usually, in the aggregate itself, smaller microstylolite seams between the individual grains.

Borders of hydromuscovite develop along the microstylolite seams repeating their intricate pattern and replacing quartz and feldspar grains. Hydromuscovite borders are often monocrystalline with the basal (001) surface of a flake oriented parallel to the outlines of the border.

The columns are most intensively replaced by hydromicas especially along the sides, where pressure was evidently quite great. Some quartz and feldspar columns are completely replaced by hydromuscovite and in such a way that the basal planes of the hydromuscovite flakes are parallel to the length of the column. Hydromuscovite developed in the grains penetrated by the columns forms festoonlike structures with the convex sides of the festoons directed towards the grain which is being replaced (Fig. 1). The micaceous films of partings developed along the seams are sometimes quite thick (as much as 2 mm). In that case they are composed of random aggregates or high interference colors and with larger (0.2 mm) palmate flakes of muscovite developing at the expense of the smaller ones.

The sericite aggregates occasionally contain muscovite flakes with relict pleochroism in the greenish tones, inclusions of rutile needles (sagenitic net) and, less commonly, minute crystals of anatase and brookite formed as a by-product of the replacement of biotite. In such seams, relict grains of quartz and feldspar of intricate shape, obviously representing the remnants of microstylolites, which have cut completely through the grains into which they have been pressed, are often found together with almost perfectly fresh detrital grains of tourmaline and zircon, corroded crystals of epidote, well-formed crystals of secondary rutile, anatase and brookite, opaque titanium oxide, commonly forming chainlike accumulations, and veinlets of magnetite.

The microstylolite surfaces commonly split into smaller surfaces, unite again, branch, and curve (Fig. 1), so that their direction roughly coincides with the lamination or is at a small angle to it. Evidently, stylolitization develops along cracks parallel to the lamination planes, covered by accumulations of flakes of biotite and clay minerals altered into sericite and hydromuscovite. The latter develops also during the process of stylolitization at the expense of quartz and feldspar. The microstylolite columns and festoons are normal to the laminations.

The intensive development of intricate mycrostylolite surfaces observed in some specimens and the associated sericitization produce distinctive mica-rich sandstones.

Microstylolitic contacts between grains are much more common. They differ from the seams by being localized, not along the definite more or less flat planes of lamination or in fractures, but in the small parts of the detrital grain surfaces which come in contact with each other and show a common boundary in thin sections. Most commonly such microstylolites form at contacts between quartz grains, less commonly at contacts between quartz and feld-

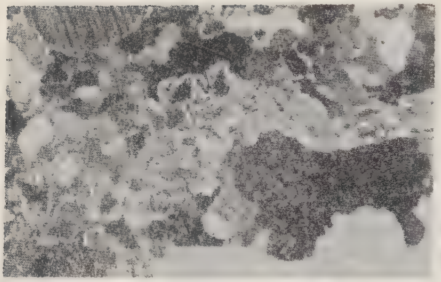

FIGURE 1. Microstylolite seam in sandstone with flakes of secondary hydromuscovite (light). Mirnoye borehole; depth, 1595.77-1600.57 m.; x 23; crossed nicols in all photo micrographs.

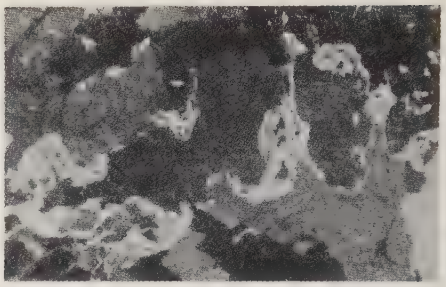

FIGURE 2. Hydromuscovite borders replacing quartz and microcline along microstylolite seams. Mirnoye borehole; depth, 1595.77-1600.57 m; x 23.

spars and between grains of microcline. Rarely microstylolites are found at contact between quartz or feldspar grains and flakes of biotite (Fig. 5).

Under the microscope these microstylolites have the appearance of somewhat asymmetrical teeth widening towards the base and penetrating into the adjacent grain.

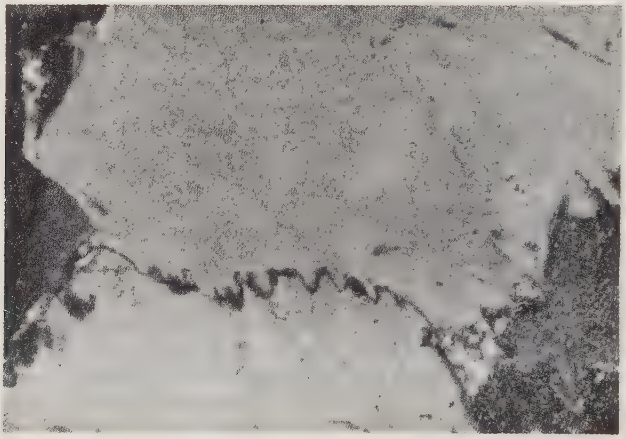

FIGURE 3. Microstylolitic structure at contact (quartz grains). The contact is bordered by hydrous iron oxides (black). Mirnoye borehole; depth, 1589.28 m; x 83.

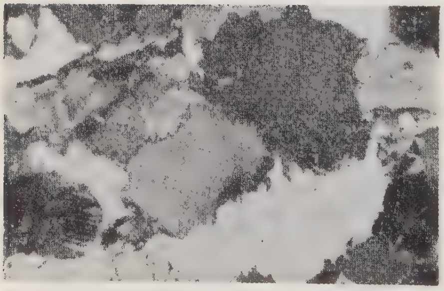

FIGURE 4. Grain boundary microstylolite. Mirnoye borehole; depth, 1589.28 m; x 23.

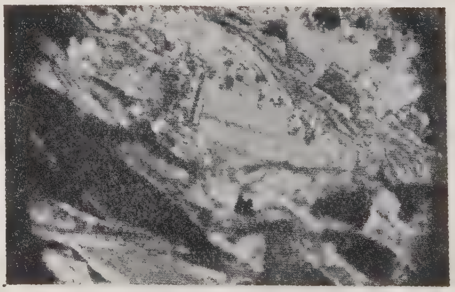

FIGURE 5. Microstylolite at the contact between a plagioclase grain and a flake of biotite. Kaushany borehole; depth, 1391-1393 m; x 23.

Their size varies within a wide range, from 0.04 to 0.5 mm in length and from 0.01 to 0.15 mm in width at the base. Sometimes the teeth have needlelike points; sometimes they terminate in a flat or slightly convex, commonly somewhat serrated surface. In some grains the teeth resemble birds' claws; in others, the teeth of a saw (Fig. 4).

The teeth are usually roughly parallel to each other and normal to the lamination. They are variously oriented with respect to the grain surfaces, giving the latter a step-like structure; the teeth are developed best when they are perpendicular to the grain surface, and are almost or completely invisible when the grain surface is nearly parallel to the direction of stylolitization or coincides with it. Stylolitization, therefore, is most clearly seen on the upper and lower parts of the grain surfaces paralleling the lamination. Besides this principal direction, the microstylolite columns sometimes follow other directions variously oriented with respect to the lamination and then they are normal to the contours of adjacent grains. Such columns are much more weakly developed and are much smaller than the columns developed in the principal direction.

Microstylolites occur only in those areas where the grains are in contact by their original surfaces. Stylolitization has not been observed at contacts between regenerated surfaces. There is commonly a parting along the stylolite seam which varies in composition from place to place. Where the original surface of the grains is coated with hydrous iron oxides, the parting is also of hydrous iron oxide (Fig. 3), whereas in the grains in which the original contour is outlined in sericite, the parting is also of sericite. Sometimes there is no parting and the grains appear as a single grain in plain transmitted light. In this case, microstylolites become visible between crossed nicols because of the difference in the orientation of the grains. The ferruginous film thickens on the flattened obtuse points of the stylolite columns (or in the depressions between them). Along the sides of the columns it thins out and disappears (Fig. 3).

The micaceous partings show no regular pattern. They may coat a column with uniform thickness or may thicken along its sides. Sometimes the columns are pseudomorphously replaced by a single hydromuscovite crystal with its base parallel to the length of the column. Both feldspar and quartz columns may be so replaced. The micaceous parting is usually thickest at contacts between grains of quartz and microline. Sometimes under the microscope minute grooves and striations may be seen on the sides of a column parallel to the length, indicating movement of grains with respect to each other.

Peculiar microstylolite forms occur at contacts between biotite and quartz or feldspar. Rarely the contact shows small columns of the usual type, 0.01 to 0.02 mm in width and about 0.04 mm in height with the biotite of the column being bleached and replaced by muscovite of hydromuscovite (Fig. 5). More commonly the biotite penetrates into quartz or feldspar in the form of a festoon with smooth outlines. These festoons are accompanied by a parting of secondary hydromuscovite which reproduces the contours of the biotite festoons. The festoons corrode the detrital grains giving them an intricate relict aspect. The convexities of the festoons point towards the detrital grain. They do not show any regular orientation with respect to the structural elements of the rock but are commonly normal to the lamination and as a result the detrital grains appear flattened in that direction.

Microstylolites with grains are rather rare. They have been found in quartz and microline grains from the sandstones of the Mirnoy borehole.

Under the microscope, these microstylolites are revealed as sinuous or serrated lines cutting across the grain. The length of the teeth does not exceed 0.1 mm. In quartz such lines are oriented alike and the parts of a grain on the two sides of the seams have a slightly different optical orientation. A thin film of hydromica (thousandth of a millimeter) develops along the seams. In microcline grains the sharply serrated microstylolite seams are oriented in different directions. Sometimes they coincide with the (001) and (010) cleavages. The microcline grids in the parts of the grain at the contact do not coincide, indicating a certain amount of solution along the seam. In some microcline grains there are signs of mechanical deformation and formation of blocks with different optical orientation and the edges of the blocks bear serrated microstylolite seams with micaceous parting (Fig. 6).

The microstylolites in grains form along fractures which sometimes coincide with cleavage cracks and sometimes form as a result of stress. The sharpest microstylolites in grains develop in the direction of lamination.

There are no substantial differences of opinion on the formation of stylolites in

sandstones. Most investigators ascribe it to the solution of detrital grains under pressure [12], which sharply increases the rate of solution in a number of substances. Indeed, the deep penetration of one grain into another in the form of columns, teeth, and knobs without any signs of mechanical or plastic deformation of the grains clearly points to solution. Evidence of considerable pressure is manifested not only by the form of the microstylolite columns but also by the presence on their sides of very fine grooves and by the character of distribution of ferruginous and micaceous partings. The same mechanism is appealed to by most investigators to explain the formation of styolites in limestones.

The variation in the form of stylolites and sutures has not been completely explained. It may be because of differences in the strength of bonding between individual areas of cementing material (hydrous oxides, clay, etc.) and the grains. The cementing film adhering tightly to part of a grain, presumably protects that part against solution. The difference in pressure at the contact between the grains, depending on the difference of contact areas and the orientation of the touching surfaces in relation to the compressive forces, must result in different rates of solution of the detrital grains. It is probable that the differential solution producing microstylolitic seams in the grains is determined also by the characteristics of fine structure of the crystals, twinning, susceptibility to plastic deformation, the presence of dislocations, and their orientation and distribution. Undoubtedly the size of the microstylolites is controlled also by the composition and temperature of solutions and the duration of the process of solution.

The microstylolite partings in sandstones, in spite of their resemblance to those in limestone stylolites, are of an entirely different origin. In limestone they consist of the insoluble materials present in it. In sandstone they form partly from the cementing material which originally coated the detrital grains and partly from the secondary minerals crystallized from solutions or developed by replacement of terrigenous material during the process of stylolitization.

Structures produced by mutual shape accommodation (conformal structures) and by penetration (incorporation structures) are as common as microstylolites.[1]

[1] The structure similar to those here called conformable and incorporation structures are described in American literature under the name concavo-convex.

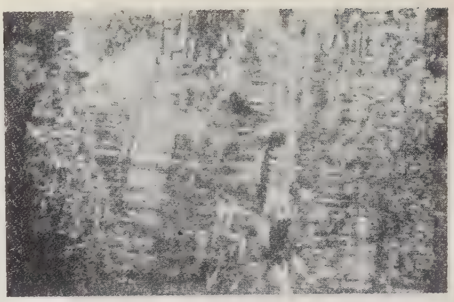

FIGURE 6. Microstylolite in a microcline grain. The seams are bordered with hydromuscovite (light). Mirnoye borehole, depth, 1595.77-1600.57 m; x 23.

Conformal structures are often observed during microscopic study in the form of aggregations of detrital grains of one and the same or different minerals without any sign of secondary growth of the grains (Figs. 7 and 8). There is a strict conformity of the contours in the aggregated grains, an absence of pore spaces between them and usually a smooth contact surface (line in sections) sometimes bearing a thin film of hydromuscovite.

The nature of the contact and the usual absence in the touching grains of any signs of plastic deformation indicate that such structures were formed not by simple mechanical crowding and packing of grains but by solution which resulted in mutual accommodation of shapes with a tendency towards closer packing of the grains, increase in the area of the contact surface, and corresponding relief of pressure.

Judging by the nature of the contacts, solution affected all minerals to the same or almost the same degree. In this respect these structures are akin to stylolites but differ from them in having smoother nonsutured contact surfaces. Sometimes these contacts pass into microstylolites.

In the incorporation of penetration structures, only one of the grains is subjected to solution and the other penetrates into it preserving its original shape (Fig. 9). Here again there is a correspondence between the form of the solution cavity and the intruding part of the grain. The contacts are usually smooth and sinuous. In sandstone whose grains are coated with ferruginous film, this film is preserved during the penetration and sharply outlines the contact surface.

Sometimes sericite flakes or borders of hydromuscovite are developed along the contact; the grains penetrate slightly (0.05 to 0.1 mm) or deeply (0.1 to 0.5 mm).

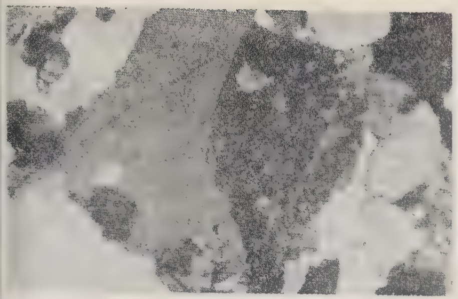

FIGURE 7. Conformal structure in an aggregate of microcline (black) and quartz (light). Kaushany borehole; depth 1604.96-1606.66 m; x 23.

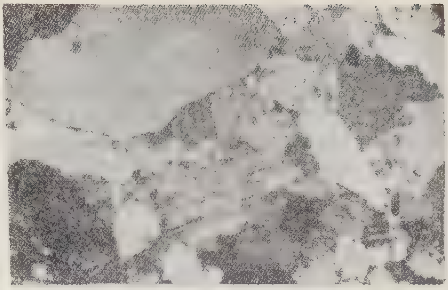

FIGURE 8. Conformal structure in an aggregate of plagioclase and quartz grains. Mirnoye borehole; depth 1604.96-1606.66 m; x 23.

ıe penetration structures occur between ɛe minerals (quartz and quartz, microıe and microcline) or between different ːinerals (quartz penetrates into feldspar, ıicrocline into quartz -- Fig. 9 -- or ıicrocline into plagioclase). Penetration of ıagioclase into microcline or quartz is ıuch less common.

Penetration structures are sometimes companied by optical anomalies in the ains indicating that they were formed unːr considerable mechanical stress. In artz or microcline penetration into quartz, e intruded grain is often granulated and e fragments form an aggregate of small riously oriented crystals.

The structures of accommodation (conrmal) and of penetration (incorporation) e widely distributed in epigenetically tered sandstones of different regions, but r a long time they attracted no attention ıd their genetic significance was ignored. ıeir intimate connection with microstyloːes and the change from one structure to ıother which can be observed in thin secɔns point to their common origin as a ısult of solution due to pressure. It is ːobable that the complex forms, characːristic of stylolites, are the result of very ıecial conditions and are, therefore, less ımmon than the conformal and incorporaɔn structures.

In a number of areas the development these structures result in the formation aggregates in which the grain shape and e character of contact resemble the ɛcrystallization textures of metamorphic ɔcks; i.e., it leads to the development crystalloblastic textures (Fig. 8).

2. Growth Structures and Secondary Minerals

The solution structures in the rocks are accompanied by regenerative growth of the detrital quartz grains and, less commonly, of feldspars.

The original boundaries of the grains become bordered by a very thin dustlike accumulation of opaque material and relicts of cement from the earlier stages of alteration consisting of hydrous iron oxides or sericite. The secondary enlargement in quartz has the same optical orientation as the detrital grain on which it grows, fills the surrounding space and cements other detrital grains. The solution and regeneration structures are very intimately related, spatially. Not uncommonly one part of the grain may be dissolved while another shows secondary growth (Fig. 9). When two adjacent grains become regenerated, euhedral crystal outlines develop at the contact of the secondary enlargements.

In some areas the secondary enlargement of quartz is such that the outlines of the detrital grains are not recognizable in

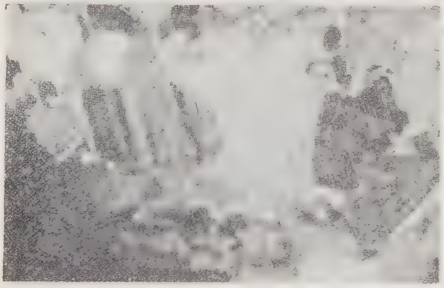

FIGURE 9. Incorporation of a microcline grain in quartz. The quartz grain is partly regenerated. Mirnoye borehole; depth, 1590.67 m; x 23.

thin sections and the area has a mosaic or granoblastic aspect.

The regeneration of recrystallized quartz grains and of the fragments of quartzite is of a different character. The random orientation of quartz particles in the regenerated grains causes random distribution of the secondary quartz in the regenerated border and as a result the latter acquires complex mosaic structure with sinuous and palmate grain outlines more or less corresponding to the original outline of the detrital grain.

The regeneration of feldspar grains occurs much less commonly than that of quartz and has been observed only in some of the beds. Plagioclase grains acquire borders of albite with the same or nearly the same optical orientation (Fig. 10).

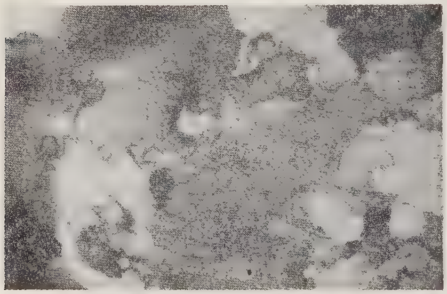

FIGURE 10. Regeneration of a detrital plagioclase grain. Mirnoye borehole; depth, 1589.87 m; x 23.

The regenerated grain is usually euhedral and somewhat elongated parallel to the vertical axis. Twinning of the detrital grain commonly appears in the secondary growth (Fig. 10). In a number of cases the boundary between the secondary enlargements of quartz and plagioclase is an irregular sinuous line characteristic of xenoblasts and indicating simultaneous crystallization of the two enlargements.

Regeneration in microcline is less common than in plagioclase. The secondary border is represented by potash feldspar without the grid and frequently with a different optical orientation than that of the original grain. The secondary potash feldspar is usually resorbed by quartz growing towards it and the outer contours of the secondary potash feldspar are serrated.

The typical recrystallization structures are commonly found together with the regeneration structures in the sandstones of the Mogilev formation. Most often recrystallization affects the fine-grained sandy material lying between the pebbles and the coarser sand particles and constituting the original cement. Such aggregates commonly contain secondary crystals of albite, flakes of muscovite, and in some cases minute crystals of anatase. The texture of the aggregates is polygonal, microgranoblastic, and when mica is abundant, microlepidoblastic. The recrystallized mass sometimes preserves relict detrital grains which escaped recrystallization. In the fine-grained and medium-grained sandstones, recrystallization affects the larger detrital material also, and as a result crystalloblastic textures are found not only in the cementing material but in whole areas of the rock as well. In places, large plagioclase crystals are recrystallized with the development within them of idiomorphic grains of albite differently oriented with respect to each other and to the enclosing crystal. More common is the recrystallization of quartz granules; these, however, are not always easily distinguished from the detrital grains with a similar structure inherited from the parent rock. Quartz grains are commonly granulated along their peripheries.

Apart from recrystallization, the effects of pressure are often manifested in the mechanical deformation of the detrital material observed almost throughout the entire section. These effects consist in the intricate arrangement of mica flakes, displacements, and bending of the twin lamellae in plagioclase, and fragmentation of grains accompanied by a certain amount of displacement of the fragments with respect to each other. Some quartz grains are mylonitized.

The secondary sericitic cement is widespread in the sandstones. In some varieties it forms the matrix for the detrital grains which may be partly regenerated; in others, it fills the interstitial spaces of some of the detrital grains. The sericite intensively corrodes the feldspar and quartz grains, mainly along the edges, penetrates into quartz in the form of thin plates, and develops along the fractures, forming distinct veinlets. In the cementing sericite mass there are also strongly corroded, relictlike quartz and feldspar grains of the silt size and smaller. In the areas where the sericite cement is developed, some of the interstitial spaces between the detrital grains contain a well-crystallized clay mineral. This mineral occurs also as a relict in the sericite mass which replaces it and is, according to all the data, a constituent part of the

cement of an earlier stage in the existence of the rocks.

The extremely strong development of muscovite among the secondary minerals should be noted. Besides occurring in borders, festoons and flakes, which form during stylolitization, it usually replaces biotite and in that case it occurs in sheaves with lens-like segregations of siderite, sagenitic nets of rutile, and minute crystals of anatase and brookite. Muscovite is found also in fanshaped, accordionlike and "stack of coins" aggregates. These structures, judging by their form, are replacements of kaolinite and kaolinitized biotite and chlorite, as shown by the presence in them of inclusions of titanium minerals and their relict pleochroism. The large sheets of muscovite with high interference colors are formed also by replacement of lamellar aggregates of sericite in the cementing mass and of the detrital feldspar and quartz (Fig. 11).

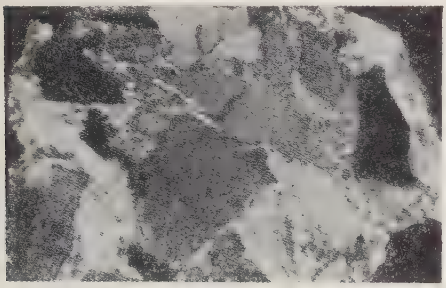

FIGURE 11. Secondary muscovite (light-colored plates). Kaushany borehole; depth, 1400 m; x 34.

The lamellar and platy muscovite is commonly found in secondary quartz and, in places, a grain of quartz is completely replaced by a flake of muscovite.

The intensive development of secondary quartz sometimes results in the replacement of muscovite, sericite aggregates, and some accessory minerals by quartz and to the development of very fine threadlike (fractions of a millimeter wide) veinlets which cut through the sandstones. The replacement of muscovite by quartz, forms distinctive pseudomorphs preserving the micaceous habit of the original crystals, their corrugations, and fanlike and sheaf-like forms. The pseudomorphs of quartz after sericite are identified not only by their form but also by the presence in them of muscovite relicts. These pseudomorphs preserve the random orientation of the sericite flakes, but the inclusions of titanium minerals and siderite characteristic of the sericite aggregates are usually removed.

Quite commonly quartz replaces tourmaline, so that only small fragments of the latter remain, with the spaces between them being filled with clear quartz. In the immediate vicinity of such replacements there is usually secondary acidular tourmaline. In places, tourmaline needles develop directly on the surface of the detrital tourmaline grains. In the sandstones of the Mirnoye borehole, quartz replaces barite, which locally appears as the cementing material in the rock.

Among other secondary minerals, the sporadic appearance of well-formed colorless octahedrons of fluorite should be noted; it is usually associated with secondary quartz.

Ankerite is abundant; it replaces potash feldspars, less commonly plagioclase, fills pores in the rock and in places resorbs secondary quartz. Judging by the relationship among the minerals, ankerite represents a later stage of epigenetic alteration of the sandstones.

III. CONDITIONS AND FACTORS OF ROCK ALTERATION

The data given above show that the rocks of the Mogilev formation are intensely altered. Besides the substantial changes in the original mineral composition and development of secondary minerals, the structure of the original rocks is also profoundly changed.

Notable among the structural modifications are the structures formed by the solution of detrital grains under one-sided pressure. The orientation of the stylolite columns indicates that the pressure was normal to lamination and since the latter is horizontal, radial with respect to the earth's surface. It is natural to assume that this pressure was due to the weight of the overlying beds. The formation of microstylolites and the genetically related conformal and incorporation structures and also the changes in the mineral composition could have occurred only after a considerable subsidence of the region and accumulation of a thick sequence of later sediments. These structures, therefore, developed a very long time after the accumulation and diagenesis of the sediments and are definitely epigenetic.

The present depth of the Mogilev formation (1500-1600 m) cannot be regarded as the maximum depth of subsidence of the

region, for there is a considerable missing interval in the section which includes Mi Middle and Upper Paleozoic and the greater part of the Mesozoic (Silurian to Cretaceous). It is not improbable that the maximum depth of subsidence of the Mogilev formation exceeded its present depth and that the overlying sediments were eroded in pre-Cretaceous time. There is no reason, however, to assume that this depth was of a different order of magnitude because the thickness of the missing stratigraphic divisions is not great in the neighboring territories and does not exceed the usual thickness of platform sediments.

M. Heald [12] noted the presence of microstylolites in the sandstones of different ages (Cambrian to Cretaceous), in the eastern parts of the United States, which lie at a depth of about 4 km (13,000 feet). He does not mention the minimum depth of occurrence of these structures. However, remarking on the presence of microstylolites in the Pennsylvanian sandstones of West Virginia, which, it is believed, were not overlain by more than a few thousand feet of sediments, Heald comes to the conclusion that stylolitization may occur at a relatively shallow depth.

G. Taylor [15] remarks on the wide development of solution structures in the Jurassic and Cretaceous polymict sandstones of Wyoming (U.S.A.) revealed by boreholes at depths from 300 to 2500 m. He proves on the basis of detailed study of the section that in the interval between 800 and 2000 m the number of convexo-concave contacts increases from 10 to 30 percent, and the number of microstylolite contacts, which appear at a depth of 1500 m, increases rapidly downwards and at the depth of 2400 m reaches 34 percent of the total number.

These data do not, unfortunately, help to establish the minimum depth of formation of these structures, since the thickness of the overlying strata overlying the Jurassic and Cretaceous sandstones is estimated differently by different investigators (from 3000 to 6400 m). Moreover, these sediments are folded, and this also may have contributed to the formation of conformal and microstylolite structures.

The present author has observed solution structures at relatively shallow depths (less than 1.5 km) in the sandstones of the Valday and Redkinskiy complexes of the central part of the Russian platform and Belorussia.

The formation of secondary minerals is intimately related to the development of solution structures. It is accompanied by the solution of a considerable amount of detrital material. This amount cannot be estimated exactly. Preliminary measurements on thin sections indicate that because of the development of the microstylolites alone, the thickness of the strata is diminished by 5 to 25 percent. This estimate, based on the length of the stylolite columns, undoubtedly gives low results, for the development of the columns is accompanied by partial solution. It is impossible to estimate the amount of material dissolved during the development of conformal and incorporation structures, since the original form of the detrital grains is unknown. The amount of fine-grained detrital material and the original cement of the sandstones which has been dissolved and recrystallized, cannot be estimated at all. The total effect of the development of solution structures suggests a value which undoubtedly exceeds the content of the secondary minerals in the rocks, which amounts to about 20 percent. Thus, the epigenetic alteration of the rocks was probably accompanied by removal of material and decrease in thickness.

The solution of the detrital quartz and feldspar grains is accompanied by precipitation of secondary minerals from the solutions and the development of complex exchange reactions between the dissolved substance and the solid constituents of the rock. Quartz is the most abundant of the secondary minerals and forms secondary enlargements and recrystallized aggregates. Less abundant are the feldspars which form rims on the detrital grains and a few recrystallized aggregates. The small amount of authigenic feldspar indicates that the bulk of the dissolved microcline goes into the making of other minerals. The alkalies react with kaolinite to form sericite; the aluminum ions participate in the replacement of feldspar by hydromica and muscovite; the alkalies and aluminum together react on quartz and, under the conditions of one-sided pressure, replace it with muscovite-like micas and muscovite; aluminum replaces iron and magnesium in biotite and changes it into muscovite, the iron in this case separating out in the form of siderite and, in association with magnesium, forming carbonates of the ankerite type. The transformation of biotite into muscovite is accompanied by separation of euhedral crystals of rutile, anatase, and brookite. The fluorine for the formation of the sparse fluorite is also probably derived from the biotite.

The character of the secondary minerals and the stability of muscovite indicate that the interstitial solutions are alkaline, whereas the presence of the carbonates indicates the presence in them of Ca ions and carbon dioxide.

Examination of thin sections has shown t the development of solution structures some areas somewhat precedes, but in neral coincides with, the regeneration of artz and feldspar and the development of ricite, hydromicas, and other secondary nerals.

Simultaneous, or almost simultaneous, ution and precipitation of quartz and feld-ars points to the saturation of the solu-ns and indicates that the system is near uilibrium. The recrystallization of the id phases is determined by thermodynamic rameters. The grains in the mechanically erstressed areas are most easily dissolved cause of high pressure and also small ains with excess surface energy. According N. A. Ogil'vi's thermodynamic computa-n [8], the solubility of the mineral's mework in the rocks depends directly on difference in pressure exerted upon it d the pressure of water within it. This ference may attain a considerable magni-e because of the enormous stresses be-een individual grains with small areas of ntact.

To the stress differences at different nts in the rocks will correspond in the ncentration of interstitial solutions and s will cause a certain amount of migra-n of material and crystallization of cor-sponding minerals in the areas where conditions favor it (growth of large ains, filling of pores, etc.

Thus, there is a direct connection between development of solution structures and regeneration of detrital grains, the velopment of secondary cement, the forma-n of crystalloblastic textures, and the mation of secondary mineral associations, ., all epigenetic transformations. The in factor in this is pressure determined the depth of burial of the sediments and temperature corresponding to this depth. doubtedly, a significant role is played also the initial composition of the solutions, ich, however, changes during the long riod of epigenetic transformation, adapting elf to the composition of the rocks.

These processes are apparently very w and for this reason the degree of inge in the rocks is directly dependent time, i. e., on the duration of the favor-le conditions.

The relation between the structures and epigenetic transformations of the rocks, apparent in the sandstones of the Mogi- formation, is evidently universal. In ier words, the development of the solu-n structures is usually accompanied by regeneration of detrital grains, formation of the secondary quartz cement, crystalloblastic structures and greater or less change in the mineral composition of the rocks. Indeed, a number of authors [12, 13, 14, 17] have remarked on the meta-quartzitic aspect of the sandstones with microstylolites and on the profound structural changes accompanying the development of secondary enlargements on quartz. These features have been observed by the present author in sandstones from different regions.

Similar profound structural changes in sandstones accompanied by the formation of a number of secondary minerals including muscovite are common, according to A. G. Kossovskaya and V. D. Shutov in the deposits of the western Verkhoyansk region [2, 3, 4].

The epigenetic processes in sandstones have well-defined limits. They continue until the pressure between grains becomes sufficiently high to cause their solution. The smaller grains gradually disappear and the pressure diminishes as the result of increasing areas of contact due to differential solution, secondary growth and filling of pores with secondary minerals.

Thus, under the platform conditions the process of recrystallization of sandstones cannot attain its end, i. e., a complete recrystallization of all of the detrital material, but gradually weakens and stops. The completeness of transformation of the rocks, besides depending on the pressure, which is determined by the depth of burial, the magnitude of the load of the overlying sediments, and time, depends also on the granulometric composition and homogeneity of the rocks, the rationbetween the detrital grains and the cementing material, the composition of the latter, the structural characteristics of the rocks, their porosity and some other physical properties. A considerable role is played, apparently, by the hydrological regime, for it controls the removal of material from the rocks. All this means that during the post-diagenetic stage of transformation of sandstones there is a specific stage characterized by an intensive change in the mineral composition and structure through the solution of the main clastic constituents, the heavy minerals in part, and by the development of secondary minerals ("cold metamorphism").

The upper limit of this stage is marked by the appearance of the signs of solution under the pressure of the main rock-forming minerals, quartz and feldspars, and of the lithic fragments (the necessary depth of burial for this is tentatively set at 1 to 1.5 km); the lower limit is set by the decrease in porosity and permeability to the values characteristic of crystalline rocks.

SUMMARY

In the deeply submerged parts of the sedimentary mantle of the Russian platform, because of the pressure produced by the overlying sediments, clastic rocks are partially recrystallized and undergo profound textural changes accompanied by the formation of a distinctive set of secondary minerals. These transformations, which are not characteristic of the normal sedimentary rocks, occur during a definite late stage of epigenesis and are identical with the changes typical of the early stages of metamorphism. These altered rocks, therefore, may be regarded as transitional between sedimentary and metamorphic.

Inasmuch as the beds of the Mogilev formation lie within the platform and are horizontal, their alteration cannot be ascribed to deformation. The alteration occurs without any noticeable additions of material from the deeper zones of the earth and results from the regrouping of the original material of the sedimentary rocks themselves.

The presence of rocks with clear signs of regional metamorphism in the deep parts of the sedimentary mantle of the platform requires a revision of the existing idea that regional metamorphism takes place at great geosynclinal depths and that magmatic influences play a leading role in it.

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THE AGE AND ORIGIN OF THE SO-CALLED "TILLITES" OF THE NORTHERN PART OF THE YENISEY RANGE

by

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The authors describe peculiar boulder-pebble argillites which are widely distributed in the northern part of the Yenisey Range and are usually called "tillites," and conclude that they are Lower Cambrian conglomerates formed at the base of an uplift (cordillera) and transported over a considerable distance by subaqueous slumping.

* * * * *

Widespread in the northern part of the Yenisey Range are peculiar coarse-grained deposits consisting of pebbles and boulders of various sizes of both sedimentary and igneous rocks imbedded in a sand-clay matrix and known in the literature as boulder-pebble argillites.

These deposits were first described from the Vorogovka and Suktal'ma Rivers by I. G. Nikolayev [12], who called them Lower Cambrian moraines. Later, N. S. Shatskiy [17] and Ye. V. Shchukina called them fanglomerates. A. N. Churakov [16] strongly defended the glacial origin of these rocks and attempted to prove that they had been deposited under water (marine tillites). He referred them to the Proterozoic. In 1954, O. P. Goryainova, E. A. Fal'kova, and G. F. Lungersgausen described outcrops of these rocks from the Chivida River (right tributary of the Chapa River) and from the Bol'shaya Chernaya River (left tributary of the Podkamennaya Tungusska River). These authors shared Churakov's opinion of their glacial origin, but Fal'kova considered them Lower Cambrian whereas her co-workers regarded them as Proterozoic. G. I. Kirichenko [8] included Vorogovka "tillite" in his Upper Proterozoic Chingasan series and, without adducing new proofs, also referred them to glacial-marine deposits.

New data on the geology of the "tillites" of the northern part of the Yenisey obtained by the present authors in 1957 during traverses along the Chapa, Suktal'ma, Chivida, and Vorogovka Rivers are given below.

The boulder-pebble argillites are a part of a complex and varied formation comprised of greenish-gray sandstones, argillites, siltstones, and subordinate dolomites. The formation is Lower Cambrian and has been described by the geologists of the All-Union Aerogeological Organization as the lower part of the Chapin (Porozhikhin) formation, but is separated by us as an independent Chivida formation.

On the Teya River, near the village of the same name, the Chivida formation is underlain, apparently conformably, by Lower Cambrian redbeds; but elsewhere it lies with a pronounced unconformity on two Upper Proterozoic formations, the Pogoryuy and the Uderey. A description of the latter and of the Lower Proterozoic Penchenga and Karpinskiy Range formations has been given by G. I. Kirichenko [7]. The Chivida formation is overlain by deposits named by us the Nemchansk formation. This is a redbed formation with subordinate dolomites in its lower part. Finally, at the top of the Lower Cambrian series lies the carbonate Lebyazh'ya formation containing in its middle part, trilobites of the Tolbachan horizon of the Lena stage [4].

The easternmost section of the Chivida formation, which contains boulder-pebble argillites, outcrops on the Chapa River near the mouth of the Chivida River (Fig. 5). The following strata lie here with strong unconformity on the Upper Proterozoic Pogoryuy formation (from the base up).

1. A group of greenish-gray argillites rhythmically alternating every 5 to 10 cm with gray siltstones and, near the top, with lenses of sandstone as well. In the lower part there are some layers of gray dolomites:

Thickness 360-380 m.

2. A sequence of argillites similar to those of group 1 and occurring in layers 0.5 to 1.5 cm thick interbedded with medium-grained, quartzitic, commonly cross-bedded sandstone beds from 0.4 to 0.8 m thick. At the base of some of the layers there are sparse well-rounded pebbles of quartz and feldspar, less commonly of chert, gray quartzite, and dark shales ranging from 0.5 to 2 cm in diameter. The lower bedding surfaces bear tongue-like and rib-like markings. The rocks show clear signs of subaqueous slumping. In the upper part of the sequence the argillites disappear. Thickness 80 m.

3. Interbedded argillites and siltstones similar to those of group 1 but dirty red in color. In the lower part there are layers of gray argillaceous dolomites. Thickness 60 m.

4. This part of the section contains the boulder-pebble argillites. The main cementing mass of the rocks is composed of dirty-gray unstratified arenaceous argillite. The sandy admixture contains grains of quartz, occasional feldspar and lithic fragments (carbonate rocks, quartzites, and shale). Dispersed through the argillites are granules, pebbles, and boulders ranging in size from fractions of a centimeter to 60 cm (commonly 2 to 5 cm). The majority of the pebbles are well rounded but the larger ones are usually better rounded than the smaller. Pebbles play a subordinate role in the deposit (usually amounting to not more than 15 percent and are irregularly distributed in it, some parts of the deposit containing almost no pebbles, others containing many. The distribution of large pebbles and boulders is even more irregular. Most commonly they are absent altogether, in places they occur as single specimens, and in places they are so concentrated as to touch one another. The pebbles are of the following types: light-gray and bluish-gray dolomites and dolomitized limestones, in places stromatolithic (these rocks predominate among the pebbles and form most of the large boulders), light-gray saccharoidal, in places banded dolomites, dark-gray and black argillaceous shales with a strong odor of hydrogen sulfide, light-gray and pinkish quartzites, dark shales and phyllites, cherts, vein quartz, and in a few places different igneous and metamorphic rocks such as gneisses, gabbros, granites, and amphibolites. The proportions of these rocks are given in Table 1.

In the upper part of the deposit there is a 50- to 70-meter bed of coarse-grained, coarsely cross-bedded sandstone with isolated lenses and layers of tightly packed pebbles of the same composition as in the argillites and measuring from 2 to 4 cm in diameter (very few 25 to 50 cm). Unlike the argillites, the sandstones contain more pebbles of terrigenous rocks and fewer of carbonate rocks. At the top of the deposit lies a 2-meter bed of coarse-grained sandstone similar to that just described and containing dispersed similar pebbles, 0.5 to 3 cm in diameter. From the underlying argillites and from the rocks of sequence 5, this deposit is separated by sharp boundaries but without any signs of unconformity. The total thickness of the deposit is 500 m.

5. This sequence is composed of 4- to 6-meter beds of fine-grained gray polymict sandstones and 1- to 2-meter groups of alternating (every 10 to 20 cm) layers of the same sandstone with layers of siltstones and argillites. The lower bedding surfaces bear various markings. The thickness of the sequence is 200 m. In its structure it is similar to the sequence described in detail by V. N. Grigor'yev [3] from the vicinity of Teya.

All sequences are conformable and pass one into the other gradually; their total thickness is 1,200 m. They are conformably overlain by the Nemchansk formation.

The strata described above are exposed on the northern limb of an anticline. The same section of the Chivida formation is exposed on its southern limb along the Chapa River also, although not so well, from the mouth of Krutyaka Creek to the mouth of the Chingasan River. To the south of the Chingasan River, near the mouth of Veselyy Creek, the outcrops of the Chivida formation are concealed under water but appear again 2 km higher along the Chapa River near the mouth of the Suktal'ma. The outcrops here are poor and the nature of the section must be judged from the material in slumps and slides. In general this section appears to be very much like that described from near the mouth of the Chivida River, but its thickness, so far as can be judged from the poor exposures, does not exceed 1,000 m. Although the thickness of the formation as a whole decreases, that of the pebble-boulder argillite increases and may be estimated at 699 m. The pebble-boulder argillites at the mouth of the Suktal'ma River are similar to those in the Chivida River outcrops. The composition of their pebbles is given in Table 1.

From the Chapa River, the sediments of the Chivida formation follow the general southeast and northwest structural trend. To the southeast from the Chapa, the number of pebbles in the formation gradually decreases. Talus and outcrops of the pebble-boulder argillites can be traced for several kilometers up the Chingasan and Chivida Rivers. In the upper courses of these streams the

Chivida formation is concealed under the younger Nemchansk formation but reappears on the Noyba and Teya Rivers in the area between Lopatinskiy Creek and the village of Suvorovskiy. Throughout this area, pebbles and boulders are absent from the Chivida formation and it is composed entirely of sandstones, siltstones, argillites and, less commonly, dolomites similar to those described from the Chivida River and usually exhibiting rhythmic stratification. The same rocks are found in the Chivida formation to the east, northeast and north from the mouth of the Chivida River; on the Teya River they outcrop near the Yukhtalik River, near Berezovyy Creek and near Talyy Island; on the Chapa River, near Tayezhnyy Creek, and in the upper part of the Bol'shaya Zhaduga River.

Thus, the mouths of the Chivida and Suktak'ma Rivers occupy the easternmost position in the region of distribution of the boulder-pebble facies of the Chivida formation (Fig. 5). From the mouth of the Suktal'ma these argillites can be traced continuously along its right bank to the mouth of the Vorogovka River. Within this distance, the boulder-pebble argillites replace the normally bedded sandstones, argillites, and siltstones of the Chivida formation to a still greater extent, for on the Yorogovka River boulders and pebbles are found throughout the entire section.

The exposures of the Chivida formation on the Vorogovka continue without interruption for 40 km, from the mouth of its first right tributary below the Olen'ya River to the traps. The structure of the boulder-pebble argillites in this area is similar to that of the Chapa River. As in the eastern area, the matrix of the rocks is a more or less arenaceous argillite, but here it is schistose and strongly altered. The argillite is dirty-gray or greenish-gray and only in places pistachio-green, brown-gray, or pale cherry-red. On the right bank of the Vorogovka River, 2,600 m below the mouth of the Listvennaya River, lenses of strongly metamorphosed tuffs, 3 to 5 cm in thickness, have been found. Pebbles and boulders of various rocks are dispersed through the argillites of the Vorogovka River just as they are on the Chapa River. Their size ranges widely and in a single outcrop all gradations may be observed from granules to boulders 1.5 to 2 m in diameter. It is noteworthy that compared with the argillites of the Chapa River those of the Vorogovka contain more abundant and larger boulders. In most cases the boulders constitute only 10 to 15 percent of the rock, but locally, for example, between 2.5 and 0.5 km below the mouth of the Listvennaya River, they are almost absent. Elsewhere, for instance 14 km below the mouth of the Bolotnaya River, the argillites, on the contrary, are tightly packed with pebbles. Still more irregular is the distribution of large pebbles and boulders. They occur only in some localities and lie singly and in groups of 3 to 5, usually with approximately the same orientation. The small pebbles show no orientation.

Pebbles 2 to 5 cm in diameter are most common. They are subrounded, not uncommonly subangular, and some are angular with only slightly smoothed edges (Fig. 1). Less common are pebbles 20 to 30 cm in diameter. They also vary in form from rather regular ellipsoidal to pear-shaped and angular, but poorly rounded fragments are relatively rare among them. Some large blocks are found among the pebbles and boulders (usually 1.5 to 2 m, but some 4 and 5 m across). The number of large blocks increases sharply towards the base of the formation near the remnants of ancient relief from which they were derived.

FIGURE 1. Boulder-pebble argillites on the Vorogovka River 1 km above the mouth of the Listvennaya River

The large blocks also vary in form from angular irregular fragments and plates to well-rounded large boulders. It should be noted that among the pebbles of all sizes specimens even with a single facet, "flat irons," typical of glacial deposits, are absent [22]. Some large pebbles are polished, but a careful search failed to discover striated pebbles. Some pebbles preserve signs of the original corroded weathered surface emphasizing various textural characteristics of the rock (algal layers, etc.).

Most of the larger pebbles and almost all boulders are composed of light-gray, bluish-gray and pinkish fine-grained, stromatolitic

Table 1

Composition of Pebbles from the Boulder-Pebble argillites of the Chivida Formation (%)

	Composition of pebbles						quartzites	shales and phyllitic shales	vein quartz	chert	igneous and metamorphic rocks
	total	lt. -colored dolomites	dark clayey limestones	green and red banded limestones	sugary dolomites	others					
Chapa R., mouth of Chivida R., No. 1183	73	46	16	5	6	--	10	2	2	3	9
2km above Suktal'ma R., No. 1146	59	not counted					20	13	8	--	--
Vorogovka R. 3km above mouth of Listvennaya R., No. 1195	74	45	14	3	11	--	19	8	--	--	--
700m above mouth of Listvennaya R., No. 1197	65	18	28	5	9	5	21	13	--	2	--
1.2km below, No. 1203	67	40	--	--	18	9	11	13	5	--	4
2.5km above Zakhrebetnaya R., No. 1211	64	46	6	--	12	--	24	8	--	--	4
4km above Bolotnaya R., No. 1222	61	33	--	3	21	4	24	10	--	--	5
4km above Bolotnaya R., No. 53	95	80	5	--	10	--	3	2	--	--	---
2.5km above the traps, No. 61	52	46	6	--	--	--	30	18	--	--	--
Arithmetic mean	67.8						18.0	9.7	1.7	0.6	2.4
Relative mean error	12.6						37.0	43.5			

dolomites, less commonly of various granites (gray, pink and red, microcrystalline, and pegmatitic), gray biotite gneiss, and rarely augen gneiss. All poorly rounded large blocks are dolomites. Among the smaller pebbles, besides the rocks listed above, there are also light-gray saccharoidal sandy dolomites, pink sandy dolomites, black and dark-gray fine-grained argillaceous limestones giving off hydrogen sulfide odor when struck, cryptocrystalline banded greenish-gray and cherry-red limestones, black and gray cherts, vein quartz, light-gray, gray and pink quartzites, quartzitic sandstones and siltstones, dark argillaceous and argillaceous-chloritic shales and phyllites, and some schists (mica, quartz-mica, and quartz-feldspar-mica) and amphibolites. Goryainova and Fal'kova have noted also a crystal tuff greenstone.

In most cases it is possible to name the formation from which the pebbles were derived. The schists undoubtedly came from the Penchenga (Osinovsk) formation; the phyllites and shales, from the Uderey; the quartzites and siltstones, from the Pogoryuy (Mt. Vysokaya) formation; and the banded limestones from the Kartochka formation. The light-colored fine-grained dolomites, which predominate among the pebbles, are like the dolomites of the Upper Proterozoic Chernorechensk formation.

Crystalline rocks occurring among the pebbles are similar to the rocks of the Karpinskiy Range and to the granites intruded into them. Volcanic greenstones, altered lavas and tuffs are found in the Uderey formation along the Kutukas River and in other localities. The rest of the rocks are not characteristic of any single formation but occur in various Precambrian formations of the Yenisey Range. No erratics have been found among the pebbles.

The quantitative distribution of pebbles of different kind is shown in Table 1.

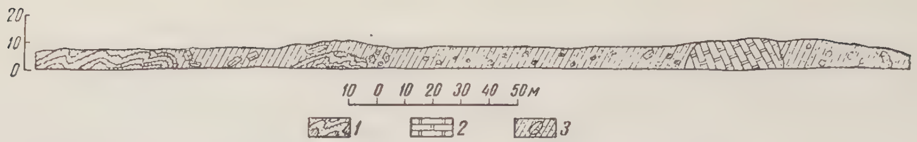

FIGURE 2. Sketch of an outcrop on the right bank of the Vorogovka River 2.5 km upstream from the traps

1 -- Outliers of Uderey quartzites and shales; 2 -- outliers of dolomites; 3 -- schistose boulder-pebble argillites of the Chivida formation.

The table shows that various carbonate rocks account for the largest number of pebbles, that quartzites hold the second place, and shales and phyllites, the third. Other rocks constitute only a few percent of the total number of pebbles. The proportion of pebbles of different rocks varies considerably.

Let us review the data on the occurrence of the boulder-pebble argillites of the Chivida formation on the Vorogovka River and their contacts with the underlying and overlying beds. The contact of the argillites with the underlying beds is exposed on the right bank of the Vorogovka River 2.5 km upstream from the place where the river enters the traps (Fig. 2). At the downstream end of the outcrop there is an exposure of the Uderey formation consisting of dark phyllites and laters of quartzite, crumpled into small asymmetrical folds. The boulder-pebble argillites lie upon it with a gently wavy contact. As shown in Figure 2, the same Uderey rocks are exposed 50 m away from the contact in the form of an outlier several meters in height and about 20 m in length which is completely covered by the Chivida argillites. Noteworthy is the fact that in this outcrop, the content of quartzite pebbles in the Chivida formation and of shales in the Uderey formation, is higher than in other outcrops (Table 1, No. 61). Another outlier with steep, almost vertical walls and measuring 30 by 40 m, outcrops 121 m upstream among the argillites. Unlike the first outlier, it is composed of massive light-colored dolomites. Large blocks of these dolomites are found in the argillites. Much larger dolomitic outliers may be seen on the Vorogovka River 4 km below the mouth of the Bolotnaya River. One of them rises 25 to 30 m above the water and can be traced for 340 m along the bank. Its base is concealed. Still another outlier has an exposed area of 60 by 100 m and barely rises above the water level. It is surrounded by talus consisting of large (as much as 6 m) angular blocks, boulders and pebbles, almost entirely of dolomite (Table 1, no. 53). In this outcrop the blocks are tightly packed and the spaces among them are filled with smaller fragments, whereas the cementing argillite is very subordinate (Fig. 3). The absence of similar large blocks near other outliers results, apparently, from the fact that the outcrops do not expose their bases, the only places where such blocks could have been deposited. These data show that the Chivida formation lies unconformably on the upper Proterozoic rocks on the Vorogovka and Chapa Rivers.

The character of stratification of the argillites can only be surmised, for the primary bedding of the arenaceous-argillaceous mass (if it ever existed) is completely masked by intensive schistosity. The strike of the schistosity is persistent along the entire Vorogovka River, varying from N 10° W to N 10° E and dipping from 50° to 70° W. The rather large deviations from this attitude which are observed in some small areas are due, apparently, to faulting.

Let us return to the outcrop 2.5 km upstream from the traps (Fig. 2). The schistosity of the argillites here cuts the contact between the Chivida and Uderey formations at an angle of 50° to 60°. This indicates with certainty that schistosity is not parallel to the bedding in the Chivida formation. In this outcrop it has a very gentle dip as indicated by the form of the base of the argillites

FIGURE 3. Boulder-pebble argillite of the Chivida formation near dolomite outliers

and the distribution of large blocks in the depressions in the underlying formation. The noncoincidence of schistosity with bedding in the Chivida formation is confirmed by the difference in orientation of the large blocks and the direction of schistosity which in many cases are at right angles to each other (Fig. 4). The attitude of groups of large blocks with like orientation suggests that the beds of the Chivida formation have a relatively gentle dip in such cases.

Direct evidence of the attitude of the Chivida beds is provided by the normally stratified sandstones and argillites which are found among the schistose rocks in two places on the Vorogovka River, apparently in the upper part of the Chivida formation. Four and a half kilometers below the mouth of the Listvennaya River and 120 m stratigraphically apart, there are two groups of fine-grained, polymict, greenish-gray sandstones of which the lower is 70 m and the upper 100 m in thickness. Six meters from the base of the upper group a 12-meter layer of strongly altered tuffaceous sandstones and tuffs has been found. The boundary between the sandstones and the boulder-pebble argillites is sharp and slightly wavy. In this outcrop the schistosity of the argillites coincides with their bedding.

At another locality, 5.5 km above the Bolotnaya River there is an outcrop of a 50- to 60-meter thick group of interbedded dark-gray and greenish-gray (sometimes with dirty cherry-red bands) sandstones similar to the ones just described and greenish-gray and dirty cherry-red argillites. These rocks form a synclinal box fold. On the northeastern limb of the syncline, towards the top of the section, the boulder-pebble argillites are gradually replaced by laminated sandstones. On the limbs of the fold, schistosity coincides with bedding (N 45° W, 55° W) but cuts it obliquely in the trough.

Considering all these data on the attitude of the Chivida formation along the Vorogovka River, it may be supposed that on the whole its beds have a gentle dip and form simple structures which are sometimes complicated by subsidiary folds and by faults. The absence of reliable measurements of the attitude of the beds and the subsidiary folding make it difficult to determine the thickness of the formation. On the basis of a number of graphic constructions, it may be estimated at several hundred meters, and it is probable that G. N. Nikolayev was right when he stated that the "tillites" are 800 m thick.

Both on the Chapa and the Vorogovka Rivers the Chivida formation is overlain by the Nemchansk formation, which is very easily identified in outcrops in spite of some

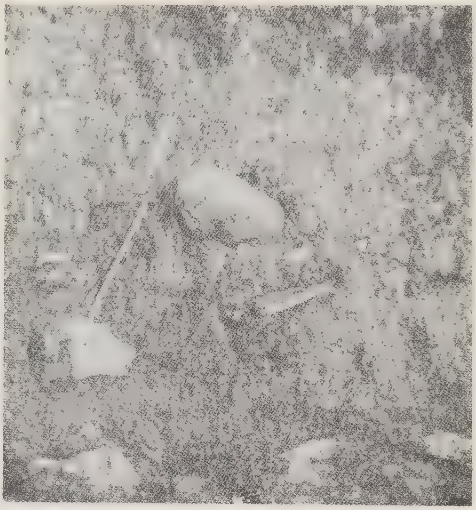

FIGURE 4. Boulders in the Chivida formation oriented normal to the schistosity. The latter is vertical in the photograph.

facies changes. That the sediments overlying the Chivida formation on the Vorogovka River belong to the Nemchansk formation also is confirmed by direct tracing of the latter. The contact between the formations can be seen on the Vorogovka River 1.2 km below the mouth of the Zakhrebetnaya River. Here the boulder-pebble argillites of the Chivida formation are overlain with a straight, sharp contact by light-gray dolomites containing thin (1 to 5 mm) partings of greenish-gray argillite and stylolite seams 1 to 3 cm in thickness. The lowermost 3 to 4 cm of the dolomites contain small angular fragments of the same rocks which form pebbles and boulders of the underlying argillites. The thickness of these dolomites is 15 to 20 m. Higher in the section they become more thick-bedded (15 to 20 cm) and rather commonly enclose layers of pinkish-yellow fine-grained arkosic sandstones.

The dolomites strike N 55° to 60° W and dip 10° to 12° E, the argillites strike N 15° W and dip 50° to 52° E, and this gives the impression of an angular unconformity. It is thus that this outcrop was interpreted by Goryainova, Fal'kova, and Kirichenko. Let us recall that these investigators referred the dolomites to the Lebyazh'ya formation of the Lena stage. I. G. Nikolayev [12], on the other hand, believed it more "correct to interpret this outcrop as an example of conformal blanketing of continental (glacial) deposits by marine sediments deposited at a later stage of the gradual transgression by a Lower Paleozoic sea." The schistosity of the "tillites" and the absence of schistosity in the dolomites, Nikolayev explained by the

"greater sensitivity of argillaceous tillites to tectonic processes." Indeed, the nearly north-south trend of the cleavage is found not only in the Chivida argillites. In a number of localities the rocks of the Uderey formation are schistose and the rocks of the Nemchansk formation show well-developed cleavage with the same trend. The degree of schistosity of the rocks, as shown by observations on the Vorogovka River near the mouth of the Zakhrebetnaya River, depends largely on their lithology. Here the boulder-pebble argillites are schistose, but in the overlying Nemchansk formation, cleavage of the same trend as in the argillites is observed only in siltstones, whereas the interbedded dolomites are not schistose. Schistosity is not everywhere developed in the boulder-pebble argillites. It is poor on the upper waters of the Suktal'ma River and absent altogether on the Chapa River. Let us recall that the schistosity of the Chivida formation does not generally coincide with bedding. We may accept Nikolayev's view and consider the outcrop described above as an example of such noncoincidence.

On the Vorogovka River the Nemchansk formation lies conformably on the boulder-pebble argillites ("tillites"). Evidently there is no erosion surface separating the formations, for otherwise the base of the Nemchansk formation would contain boulders and pebbles from the Chivida argillites, which is not the case.

To the northeast of the upper waters of the Vorogovka River the Chivida formation is concealed under the Nemchansk formation but reappears in the upper waters of the Bol'shaya Chernaya River and in the swampy watershed between the Yenisey and Podkamennaya Tungusska Rivers. According to the data of Fal'kova, Petrov, and Shenkman, mainly greenish-gray argillites and siltstones are exposed in this region. Boulder-pebble beds are not mentioned, but it is possible that they passed unnoticed because of the extremely poor exposures. Somewhat farther north, on the Bol'shaya Chernaya River below the mouth of the Losinaya River, these investigators noted, at the base of the strata corresponding to the Chivida formation, a 150 m deposit of coarse pebble conglomerate with granite and gneiss pebbles. Along the strike the conglomerate passes into small pebble and granule conglomerate with pebbles of metamorphic and sedimentary rocks and is overlain by gray sandstones, siltstones, and subordinate limestones. They did not mention coarse-grained rocks to the north and west from the Losinaya River in the analogues of the Chivida formation.

The extreme western outcrops of the boulder-pebble argillites are found on the Vorogovka River 1 to 1.5 km above the place where the river enters into the traps from the east. Near the western end of the traps, the Chivida formation is absent from the section and the Nemchansk formation lies with an angular unconformity directly on the Lower Proterozoic Penchenga formation. The Nemchansk formation consists here of light-gray, pink, and less commonly, cherry-red arenaceous dolomites with partings and layers of cherry-red sandstones and siltstones.

Farther to the west, along the Vorogovka River at the mouth of the Malaya Severnaya River, the analogues of the Chivida formation reappear in the section. They are gray arenaceous limestones with layers of greenish-gray polymict calcareous sandstones. Their thickness increases rapidly from 500 m at the Malaya Severnaya River to 4,000 m at the mouth of the Vorogovka River. These sediments were described by Nikolayev [12] as the lower part of the Vorogovka formation and by Goryainova and Fal'kova[1] as the Vorogovka group. Near the mouth of the Malaya Severnaya River these sediments are covered by the analogues of the Nemchansk Formation. From the lower course of the Vorogovka River the outcrops of the Vorogovka formation stretch southwest, with interruptions, along the western slope of the Yenisey Range. At its base a thin bed of basal conglomerates is found in places, but rudites are absent in the upper layers of the formation.

Thus, the boulder-pebble argillites of the Chivida formation are known over an area of 80 by 80 km which includes the upper course of the Bol'shaya Chernaya and Vorogovka Rivers, the basin of the Suktal'ma River and the lower reaches of the Chivida and Chingasan Rivers (Fig. 5). Our data have fully confirmed Nikolayev's conclusions concerning the facies change in the Vorogovka "tillites" to the northeast into greenish-gray polymict siltstones and sandstones (graywackes). On the Chapa River these rocks lie at the base of a conformable sequence of sediments separated by a profound unconformity from the Upper Proterozoic strata and in their upper layers contain trilobites of the Tolbachan zone of the Lower Cambrian Lena stage. These data allow us to follow other investigators and refer the Chivida formation to the Lower Cambrian Aldan stage. The boulder-pebble argillites ("tillites") of the

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[1] A description of the Vorogovka formation and the proof of its contemporaneity with the Chivida formation will be given in a separate paper. A discussion of this question here would take us too far away from our theme.

Chapa and Vorogovka Rivers belong, therefore, to the Aldan stage.

Passing now to the problem of the origin of these argillites, we must note that the geologists who considered them as glacial deposits did so on the basis of their general resemblance to a moraine. They found confirmation of this view in Nikolayev's description of two boulders with glacial polish and striae.

This evidence, however, is not sufficient to identify the rocks as tillites. Features similar to those of tillites -- absence of sorting and rounding, heterogeneity, and lack of stratification -- are found in deposits of other genetic types, such as those made by slides and slumps both on land and under water, in fanglomerates, mudflows, etc. [5, 10, 14, 18, 20]. The polish and striations on the boulders are also not necessarily connected with glacial action. Striated boulders may be produced by landslides and faults, as has been emphasized by Dunbar [21], Twenhofel [14] and Blackwelder [19]. "A boulder may be polished by water or wind and in both processes the main role is played by fine suspended material driven by stream, sea, or wind." [5, p. 278].

Thus the morphological resemblance of unsorted deposits to a moraine cannot be regarded as sufficient evidence for considering them of glacial origin. The most important criteria for determining the origin of tillite-like rocks, in our opinion, are their facies relationships and their position in ancient structures.

An important role in fixing the view that the Vorogovka boulder-pebble argillites are of glacial origin was played by the idea which became deeply rooted after the publication of A. N. Churakov's paper [16] that, as elsewhere on the globe, Proterozoic glaciation was regionally developed in southwestern Siberia. Churakov and other geologists after him referred all coarse unsorted conglomerates found in the ancient formations of southwestern Siberia to the Proterozoic, commonly without proof, and then automatically regarded them as tillites. Precisely this happened with the boulder-pebble argillites. Churakov, using only Nikolayev's material and without considering Nikolayev's convincing arguments that the "tillites" were Cambrian, referred them to the Proterozoic and without additional evidence accepted their glacial origin.

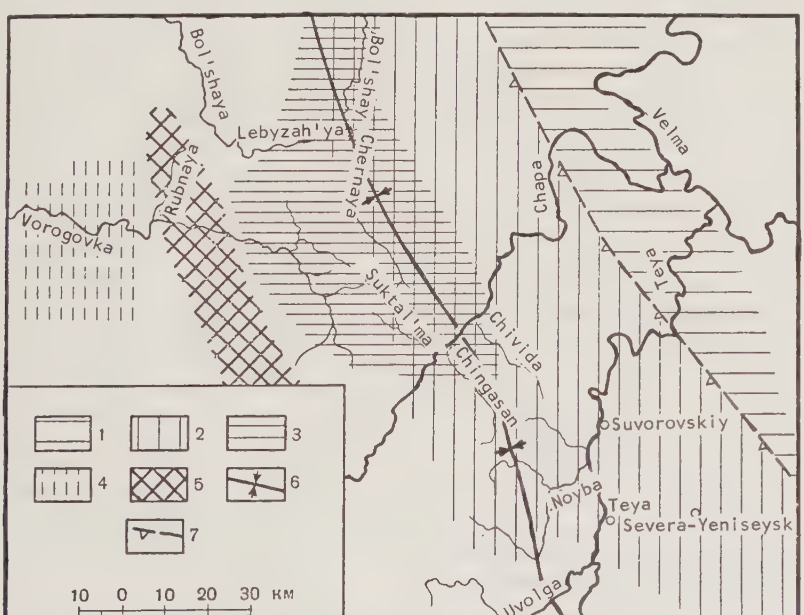

FIGURE 5. Distribution of the boulder-pebble argillites of the Chivida and contemporary formations

1 -- Thin, essentially carbonate sedimentary rocks of the Chivida formation; 2 -- thick arenaceous-argillaceous sediments of the Chivida formation; 3 -- boulder-pebble argillites of the Chivida formation; 4 -- thick carbonate-terrigenous deposits of the Vorogovka formation; 5 -- supposed zone of uplift; 6 -- axis of Teya downwarp; 7 -- boundary of the Siberian platform in the lower Cambrian time (Aldan age).

There are, however, data which argue against the glacial origin of the argillites. The argillites cannot be the products of continental glaciation, which usually affects large areas. In our case the relatively small area of the boulder-pebble argillites is surrounded by an extensive territory covered by contemporaneous marine sediments which do not contain tillite-like rocks. The local derivation of the pebbles and the absence among them of erratics also argues against connecting them with continental glaciation. The argument becomes even more convincing if we recall that the climate of this epoch in Siberia was warm, as testified by the development on the Siberian platform, of dolomites and even of gypsum and rock salt in the formations contemporaneous with the Chivida argillites and overlying them.

The argillites cannot be connected with alpine glaciation, either. Small valley glaciers could not have deposited a moraine 800 m thick over an area of 40 by 80 m. Large mountain glaciers, especially in a warm climate, could have formed only on high mountains. We cannot accept the hypothesis that such mountains existed here. The distances between the outcrops of the Aldan stage on the southwestern and northeastern limbs of the Karpinskiy Range anticlinorium do not exceed 50 to 60 km, and it is difficult to imagine a high mountain range only 50 km wide at the base. If such a range did exist in the Lower Cambrian, it would have undoubtedly left a very broad belt of coarse alluvial apron deposits along each of its sides, but such deposits do not actually exist. Moreover, the existence of such a range does not agree with the latest data of Kirichenko [7], confirmed by all the data in our possession, which indicate that in the Lower Cambrian there were no extensive tracts of emergent land in the region of the present Yenisey Range.

These arguments hold not only for the terminal moraine deposits but for fluvioglacial, recessional moraine, and other glacial deposits. The boulders and pebbles of the Chivida formation could not have been brought in by ice floes and icebergs, as was supposed by Churakov [16], for erratics are absent among them and they were obviously derived from the underlying rocks.

Clearly, the belief that the boulder-pebble argillites of the Chivida formation are of glacial origin meets with insuperable difficulties and contradicts the existing data.

The mode of formation of boulder-pebble argillites (pebble mudstones) has been widely discussed in foreign literature, and J. Crowell [20], in particular, discussed the possibility of their deposition by turbidity currents. Using the Mesozoic deposits of California as an example, he came to the conclusion that if: 1) coarse fragments are delivered to the sediments, 2) sufficiently thick masses of clay are accumulated and 3) the basin bottom has the necessary slope, then slumping will occur and the accumulated heterogeneous materials will mix to form a boulder-pebble argillite. Crowell pointed out that without sufficiently detailed study such pebble argillites are frequently mistaken for tillites.

Approximately the same mechanism of formation of pebble argillites is mentioned by Ackermann [18], who cites the example of subaqueous slumping along the shores of modern Scandinavian fiords. A similar picture of the formation of boulder and block-bearing argillites in modern sediments is given by D. V. Nalivkin [10, 11], who also mentions that old deposits of this type are often mistaken for glacial deposits.

Before giving our views on the origin of the boulder-pebble argillites of the Yenisey Range, let us review their position in the Lower Cambrian series. For this purpose we shall cite the data on the changes in the thickness of the Chivida formation and its analogues in the northern part of the Yenisey Range. From the upper waters of the Vorogovka River, where the thickness of the formation is apparently from 600 to 800 m, it increased gradually to the northeast and east approximately to the line joining the Kurepa trading post with the upper waters of the Bol'shoy Kolonok River (right tributary of the Bol'shaya Chernaya River). Thus along the Chapa River at the mouth of the Suktal'ma River the thickness of the formation is 1,000 m, at the mouth of the Chivida River it is 1,200 m, and along the Teya River near the Kurepa trading post, 1,500 m. Let us recall that with the increase in thickness of the formation as a whole, the thickness of the boulder-pebble argillites decreases in the same direction until, finally, they wedge out altogether. To the northeast from this line the thickness of the Chivida formation decreases to 800 m on the Teya River at the village of Teya, to 600 m along the Chapa River at Tayezhnyy Creek and to 230 m along the Teya River near Talyy Island.

To the west of the upper waters of the Vorogovka River lies a belt in which the Chivida formation is absent, and the Lower Cambrian Nemchansk formation lies directly on the Lower Proterozoic strata. Still more to the west, the age analogues of the Chivida formation appear once more in the section and their thickness increases rapidly (from 500 to 4,000 m) to the west.

These data show that during the Chivida time an uplift existed to the west of the

orogovka River separating two actively sub-
iding belts which may be named the Pri-
enisey and Teya downwarps. The zone of
plift lay near the crest of the present Kar-
nskiy Range anticlinorium and extended
pproximately from the western end of the
orogovka River traps to the sources of that
ver (Fig. 5). The western boundary of the
riyenisey downwarp is unknown, the north-
astern boundary of the Teya downwarp was
e relatively high edge of the Lower Cam-
rian Siberian platform.

The boulder-pebble argillites were depos-
ed at the foot of the uplift and represent a
pecial facies of normal marine sediments
the transgressive Chivida formation. Ex-
mples of similar uplifts are widely known
the literature. All geologists agree now
at they were chains of rocky islands and
ibmarine ranges or cordilleras which sup-
ied sediments to the bordering downwarps.
ne distribution of facies during the Chivida
me as described above (Fig. 5) shows that
e detrital material which contributed to the
rmation of the boulder-pebble argillites
ime from the southwest, as Nikolayev sup-
osed [12], and not from the northeast, as
elieved by Churakov [16].

All available data point to the uplift as
e source of detritus. The outcrops existing
this region reveal strongly metamorphosed
ower Proterozoic rocks of the Karpinskiy
ange and Penchenga formations and granites
truded into them. The outcrops of granite
earest to the area of distribution of the
oulder-pebble argillites lie 15 to 17 km
vay. The limbs of the Karpinskiy Range
ticlinorium have outcrops of shales and
iyllites with layers of quartzites and groups
light-colored dolomites of the Uderey and
goryuy formations. The younger Precam-
rian deposits are not known in the outcrops
thin the Karpinskiy Range anticlinorium,
though there is no doubt that they existed
the Lower Cambrian time.

The sharp variations in the composition of
e fragments depending on the nature of
iderlying rocks and the presence among
em of large blocks suggest that in many
ses transportation was short. The outcrops
the Vorogovka River 2.5 km upstream
om the traps and 4 km below the mouth of
e Bolotnaya River are especially notable in
is respect. In the second of these locali-
es the boulder-pebble argillites lie on dolo-
ites and are particularly rich in dolomite
agments; in the first locality (Table 1, no.
), they lie on dark shales and phyllites
th quartzite layers, and pebbles of these
ocks abound in the argillites.

Still another characteristic of the struc-
re of the Chivida and Vorogovka formations
is the presence in them of numerous slump marks. These marks are complex, usually disharmonic looplike folds which include either an individual layer within the bed of sandstone or siltstone or the entire bed. The thickness of layers with disturbed stratification ranges usually from 0.2 to 0.3 m. The slump marks occur in the normally stratified beds and are separated from them by sharp boundaries. Locally in the siltstones and argillites of the Chivida formation there are ellipsoidal, egg-shaped and in places very slightly rounded angular autoliths. According to A. D. Arkhangel'skiy [1], such structures are very characteristic of subaqueous slumps. It is important to emphasize that the marks of subaqueous slumping in the Chivida formation are found only to the southwest of the mouth of the Chivida River and are absent on the Teya River.

These data indicate that the Teya and Priyenisey downwarps, as well as the uplift which separated them, has specific geomorphological features and that rather steep slopes existed on the bottoms of the downwarps.

In the light of everything that has been said, the deposition of the Chivida formation may be pictured in the following way: in the Teya downwarp, bordered on one side by a relatively high edge of the Lower Cambrian Siberian platform and on the other by a rocky uplift or cordillera, rising above water in some places and submerged in others, sediments of the flysch type were deposited. In the immediate vicinity of the cordillera the products of its disintegration contributed to the formation of normal marine conglomerates which enriched the sandy-clayey sediments not only with pebbles but also with boulders and large blocks. These fragments were carried into the basin by subaerial and subaqueous slides activated by the intensive tectonism of the entire zone.

Further transportation of coarse material from the slopes of the cordillera to the deeper parts of the basin was performed by subaqueous slumping. The main arenaceous-argillaceous mass containing much colloidal material was sufficiently thixotropic so that the deposition of pebbles on its surface and other processes [18, 20] could have decreased its viscosity and caused it to move downslope as a single heterogeneous flow. Upon coming to rest, such a flow would soon become stable and the coarse material suspended in it would not settle out but remain dispersed through the clayey matrix. Sediments of this character are known in the literature as wild flysch and are considered typical of the slopes of large cordilleras.

The extensive area of distribution of the

boulder-pebble rocks of the Chivida formation and their great thickness do not contradict the hypothesis of movement by subaqueous slumping. Slumps affecting very large areas and moving enormous amounts of material are known among both modern and ancient deposits [1, 2, 6, 9, 10, 23]; some of them can be traced for tens and even hundreds of kilometers along the strike and for several tens of kilometers into the basin.

In summary, let us repeat that according to our data the boulder-pebble argillites widely developed in the northern part of the Yenisey Range and formerly described as Upper Proterozoic tillites are normal marine conglomerates of Lower Cambrian age accumulated at the bases of an uplift and transported over a considerable distance by submarine slumping. Similar unsorted conglomerates with argillaceous or arenaceous matrix are widely distributed in the ancient formations of Siberia (western part of Eastern Sayan, Baikal-Patomskoye upland) and are usually described as tillites. The investigations of M. A. Semikhatov and V. V. Khomentovskiy [13, 15] in the western part of the Eastern Sayan showed that some of these "tillites" (the so-called marine tillite on the Tyubil' River) like the "tillites" of the Yenisey Range are actually conglomerates lying at the base of the transgressive Cambrian series and accumulated at the foot of an uplift separating two Lower Cambrian downwarps. It may be supposed that other "tillites" in the ancient formations of Siberia are also piedmont conglomerates similar to the ones described in this paper.

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THE SILURO-DEVONIAN BOUNDARY IN THE NORTHEASTERN BALKHASH REGION

by

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The paper presents new data on the stratigraphy of the Upper Silurian and Lower Devonian deposits of the northeastern Balkhash region. The study of sections and faunas of these deposits shows that Upper Ludlovian, Gedinnian, and Coblenzian faunal assemblages are present and that the Upper Silurian strata of the region pass without interruption into the lithologically similar Lower Devonian beds.

* * * * *

The Upper Silurian and Lower Devonian strata are widespread in the northeastern Balkhash region and the study of their fossil faunas and floras indicates that they represent the Upper Silurian stage and both Lower Devonian stages [1, 4]. In spite of the abundance of fossils in the Silurian and Devonian sediments, the boundary between them has not been defined until now.

As far back as 1945, N. L. Bublichenko stated that "the Silurian beds pass into fossiliferous Lower Devonian beds without any observable transgressive interruption" [1, p. 49]; he tentatively separated the Gedinnian stage, but without any paleontological documentation.

Geologic mapping of the region and detailed studies of individual areas have been made between 1952 and 1957 by the Southern Kazakhstan Geological Administration. As a result of this work, additional data have been obtained, and it is now possible to refine our ideas concerning the relationship between the Silurian and Devonian deposits and to prove the presence of the Silurian Ludlovian stage and the Lower Devonian Gedinnian stage by abundant paleontological evidence.

All work was based on aerial photographs, thus eliminating possible errors in the construction of lithologically similar sequences.

The Silurian and Devonian deposits represented by lithologically similar tuffaceous sediments enter into the same geological structures broken into blocks by numerous faults. The complete section of these sediments in the northeastern Balkhash region is as follows.[1]

The Upper Silurian sediments, in places containing basal conglomerates, lie disconformably on the Upper Ordovician volcanic-sedimentary (Dzhamanshuruk) formation.

The essentially sedimentary Silurian sequence is composed of green and dark red fine-grained sandstones and siltstones in its lower part and of green polymict sandstones, sandy tuffs, thinly laminated siltstones, and some beds of silty tuffs, tuffs, and lenses of limestones and calcareous santstones in its upper part. The total thickness of the Silurian beds ranges from 1, 500 to 2, 000 m.

This Silurian sequence was subdivided by Bublichenko [1] into Llandoverian, Wenlockian, and Ludlovian stages. In 1956 the geologists of the Southern Kazakhstan Geologic Administration working in the Turanga-Saya region (Pl. L-43-48) discovered in the dark red sandstones and siltstones of the lower part of the Silurian deposits (previously considered Llandoverian) limestone beds carrying a fauna with Lissatrypa aff. linguata Buch., Plectatrypa ex gr. marginalis (Dalm.), Eospirifer nobilis (Barr.) and E. exsul (Barr.) of the Ludlovian age. In the upper part of

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[1] The Silurian and Lower Devonian faunas were determined as follows: bryozoa, by R.S. Yeltysheva; tabulate corals, by O.P. Kovalevskiy and N.V. Poltavtseva; rugosa corals, by T.V. Nikolayeva; Devonian and Silurian brachiopods, by L.I. Kaplun and T.B. Rukavshinikova; and trilobites by Z.A. Maksimova; the flora was determined by M.A. Senkevich.

ɜ Silurian sequence numerous upper Ludlo-
ın fossils were found in all localities (they
e listed in Fig. 1). Therefore, the entire
urian section is now considered Ludlovian.

M. A. Borisyak and O. P. Kovalevskiy of
ɜ All Union Geological Institute (VSEGEI)
ıdied Silurian sections and faunas of our
gion in the Kotan-Bulak Mountains and in
ɜ Tokrau-Kenterlau Valley, and on the basis
a fauna of brachiopods, tabulate corals,
d Heliolites established the age of these
ctions, formerly referred to the Wenlockian
e, as upper Ludlovian [5].

The Silurian sediments are conformably
erlain by tuffaceous-sedimentary Lower
vonian beds composed of green, tobacco-
een and red-brown polymict sandstones,
ndy tuffs, tuffs deposited in water and sub-
rially, and carbonate rocks. Siltstones are
ually absent in the Devonian section.

Fossils are abundant throughout the Lower
vonian section, and this permits the sub-
vision of the section into Gedinnian and
blenzian stages. The sediments of the
blenzian stage are more varied lithologic-
y and richer in fossils than those of the
derlying strata and contain more coarse-
ained and carbonate rocks with a greater
nge of color. The thickness of the Gedin-
ın stage ranges from 300 to 600 m and of
ɜ Coblenzian, from 350 to 650 m.

Two characteristic Silurian and Lower
vonian sections of the region are described
low.

HE KOTAN-BULAK MOUNTAIN SECTIONS (FIGURES 1 AND 2)[1] SHEET L-43-8)

S_2^{1d} 1. Greenish-gray fine- and medium-
ained sandstones and siltstones with plant
mains of the Psilophytaceae family, 50 m.

2. Green and gray-green polymict in-
uigranular sandstones containing a fauna
th Cheirurus quenstedti Barr., Dalmanites
gans Max. nom. mns. and a flora with
eniocrada (?) sp., 100 m.

3. Grayish-green polymict inequigranu-
r sandstones with occasional beds of coarse
ndstones and bluish-green siltstones. The
ndstones carry Isorthis szajnochai Kozl.,
irifer sp., Camarotoechia sp., Atrypa ex
. reticularis Linn., and Lobopyge sp.,
5 m.

4. Greenish-gray organic limestones with Lioclema sp., 25 m.

5. Green polymict inequigranular sandstones, 75 m.

6. Greenish-gray calcareous sandstones with Semicoscineum sp., Fenestella sp., Isorthis sp., Delthyris sp., Leptostrophia cf. sera Bubl., Gypidula sp., Strophonella sp., Dalmanites (Odontochile ?) sp. Max. sp. nov., and Favosites (?) borissiake Tchern., 2 m.

7. Green polymict inequigranular sandstones with rare layers of siltstones, 108 m.

8. Greenish-gray calcareous sandstones with a rich fauna containing Parmorthis sp., Isorthis szajnochai Kozl., Anastrophia sp., Stropheodonta corrugata (Con.), Bilobites bilobus L., Sieberella roemeri H. et Cl., Delthyris cf. Kazachstanica Boris., Eospirifer togatus (Barr.), Dalmanites elegans Max. nom. mns., D. (Odontochile ?) sp. Max. sp. nov., D. nominalis Max. nom. mns., 2 m.

D_1^2 9. Greenish-gray polymict inequigranular sandstones with the following fauna in the upper part: Pentagonopentagonalis subpennatus Jelt., Decacrinus equilobatus Jelt., Leptostrophia cf. sera Bubl., Leptaena ex gr. rhomboidalis W., Atrypa ex gr. reticularis L., Anoplotheca (?) sp., Pholidostrophia sp., Spirifer sp., Zosterophyllum sp., and plant remains of the Psilophytaceae family, 127 m.

10. Green polymict sandstones, 50 m.

11. Bluish-gray ash, 4 m.

12. Green polymict sandstones with Leptostrophia cf. sera Bubl., Leptaena ex gr. rhomboidalis W., Atrypa ex gr. reticularis L., Gypidula ex gr. galeata (Dalm.), Howellella mercuri (Gos) kasachstanika Kaplun subsp. nom mns., Zosterophyllum sp., and plant remains of the Psilophytaceae family, 60 m.

13. Greenish-gray, tobacco-green sandstones, sandy tuffs, and tuffs, 122 m.

14. Tobacco-green sandy tuffs, 40 m.

D_1^2 15. Brown fine-grained sandy tuffs and quartz albitophyre tuffs with abundant fossils of Drepanophycus spinaeformis Goepp., Protolepidodrendron sp. and plant remains of the Psilophytaceae family, 3 m.

16. Tobacco-green tuffs and inequigranular sandstones with the Drepanophycus spinaeformis G. fauna, 96 m.

17. Light-gray fine-grained calcareous

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[1] Section constructed by L.I. Kaplun and M.A.
nkevich.

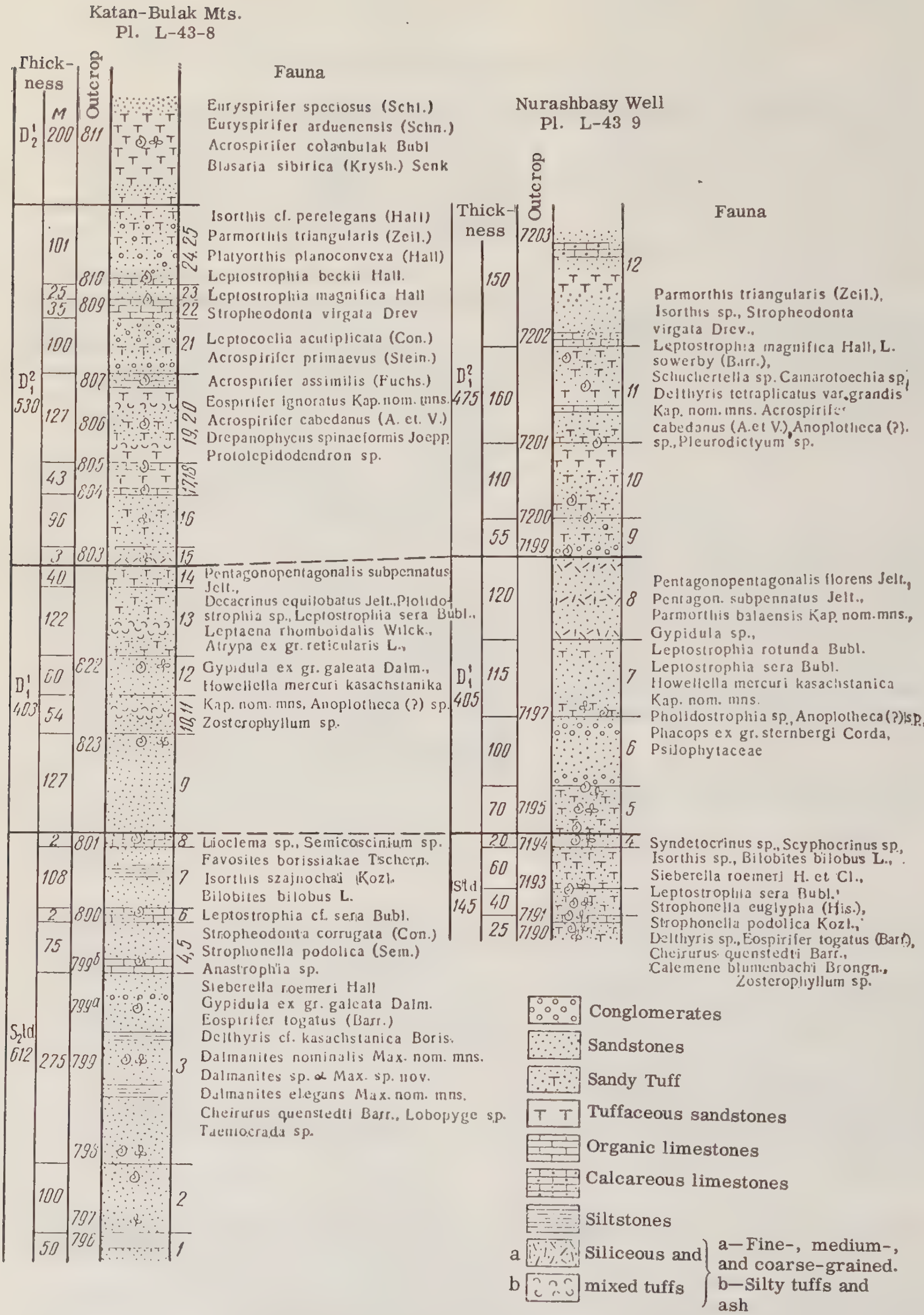


FIGURE 1. Upper Silurian and Lower Devonian sections of the northeastern Balkhash region.

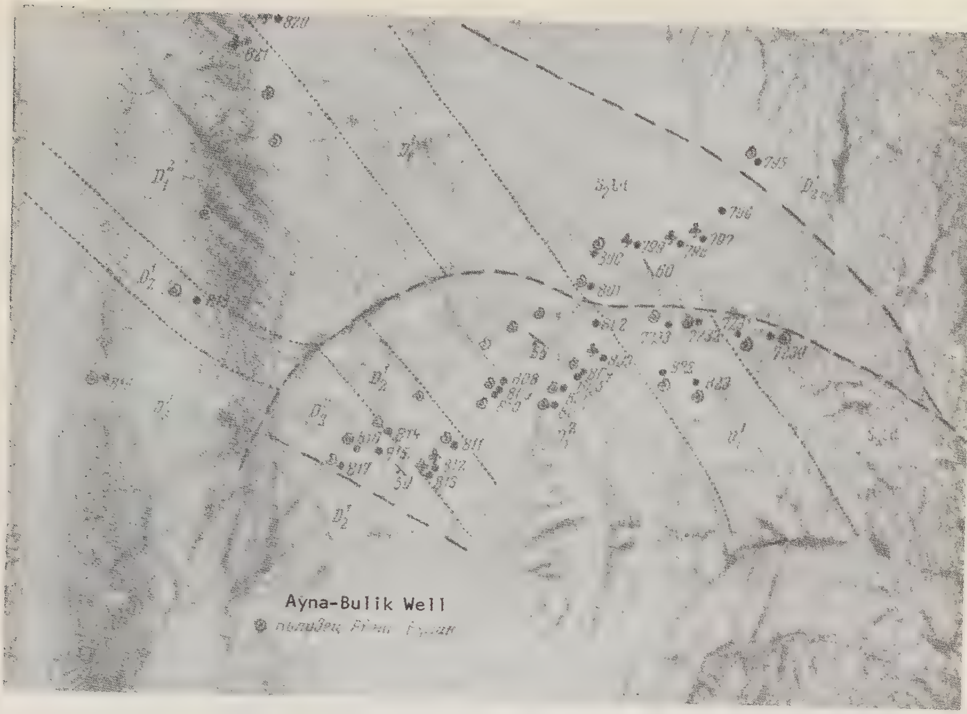


FIGURE 2. Outcroppings of Silurian and Devonian deposits in the Kotan-Bulak Mountains (Pl L-43-8).

Aerial photograph 1:30,000; fauna and flora localities are marked. D_1^1 -- different stages of the Devonian; S_2^{ld} -- Ludlovian stage of the Silurian.

andstones with Leptostrophia magnifica Iall, Eospirifer ignoratus Kaplun nom. mns., m.

18. Tobacco-green, gray-green tuffs, andy tuffs, and calcareous sandstones with bundant Isorthis cf. perelegans (Hall), eptocoelia acutiplicata (Con.) and Acropirifer cabedanus (A. et V.), 35 m.

19. Gray-brown dirty tobacco-colored, ark-gray and tobacco-colored sandstones, andy tuffs, and tuffs with Parmorthis tringularis (Zeil.), Platyorthis planoconvexa Hall) and Leptocoelia acutiplicata (Con.), 5 m.

20. Tobacco-green, dirty-green sandtones, sandy tuffs, tuffs, and limestones. eptocoelia acutiplicata (Con.) in the limetones, 62 m.

21. Gray-green, tobacco-gray sandy uffs, sandstones and conglomeratic sandtones, 100 m.

22. Light- and dark-gray limestones nd calcareous sandstones with abundant entagonocyclicus decilobatus Jelt., Kusbassocrinus kaplunae Jelt. sp. nov., Chonetes grandis Bubl., Leptocoelia acutiplicata (Con.), Leptostrophia beckii Hall, L. magnifica Hall, L. explanata (Sow.), Stropheodonta virgata Drev., S. ampliata Bubl., Acrospirifer primaevus (Stein.), A. assimilis (Fuchs), 35 m.

23. Tobacco and tobacco-green inequigranular sandstones, 25 m.

24. Gray-green calcareous sandstones with Crytina heteroclita Defr., Leptostrophia magnifica Hall, Stropheodonta bella Bubl., Acrospirifer assimilis (Fuchs), Euryspirifer cf. arduenensis (Schn.), Meristella sp., 1 m.

25. Gray-green, tobacco-green, dirty-tobacco colored sandstones, sandy tuffs and tuffaceous sandstones, 100 m.

26. Dirty tobacco-colored, reddish-brown fine-grained sandy tuffs and tuffaceous sandstones with Euryspirifer speciosus (Schl.), E. ardunensis (Schn.), and Acrospirifer cotanbulak Bubl. fauna and Blasaria sibirica (Krysh.) Senk. flora in the upper part, 100 m.

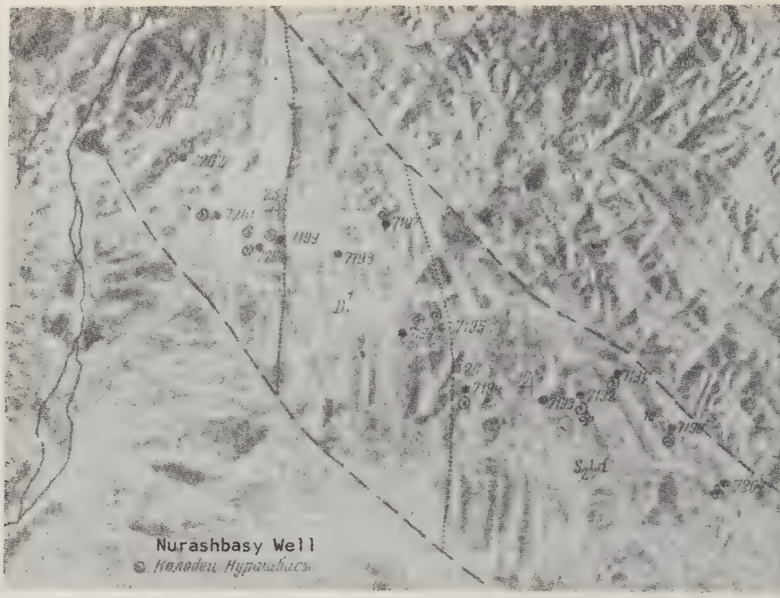


FIGURE 3. Outcroppings of Silurian and Devonian rocks in the region of the Nurashbasy Well (Pl. L-43-9).

Aerial photograph, 1:30,000, fauna and flora localities are marked. D_1^1 -- D_1^2 -- different Devonian stages; S_2^{1d} -- Ludlovian stage of the Silurian.

SECTION IN THE REGION OF THE NURASHBASY WELL (FIGURES 1 AND 3) (PL. L-43-9)[1]

S_2^{1d} 1. Tobacco-green fine-grained platy sandy tuffs with layers of calcareous sandstones filled with fossils of Pleurodictium sp., Isorthis sp., Sieberella roemeri H. et Cl., Strophonella sp., Leptostrophia sera Bubl., Eospirifer togatus (Barr.), 25 m.

2. Gray-green fine-grained and very fine-grained sandy tuffs with thin beds of coarse-grained sandstones. The very fine-grained varieties contain a varied fauna: Syndetocrinus sp., Scyphocrinus sp., Isorthis sp., Bilobites bilobus L., Sieberella roemeri H. et Cl., Leptostrophia sera Bubl., Strophonella podolica (Sem.), Eospirifer togatus (Barr.), Delthyris sp., Anoplotheca sp., Cheirurus quenstedti Barr., Calymene blumenbachi Brongn., and a flora with Zosterophyllum sp. and remains of Psilophytaceae, 40 m.

3. Tobacco-green and greenish-gray very fine-grained sandy tuffs with poorly preserved fauna and flora containing Sieberella (?) and Chonetes sp.

4. Green and gray-green very fine-grained sandstones with Syndetocrinus sp., Isorthis sp., Bilobites bilobus L., Sieberella roemeri H. et Cl., Leptostrophia sera Bubl., Strophonella euglypha (His.), Anoplotheca sp., Dalmanites nominalis Max. nom. mns., Phacops rubidus Wdkd. and plants of the Psilophytaceae family, 20 m.

D_1^1 5. Greenish-gray fine- to medium-grained sandy tuffs with abundant: Pentagonopentagonalis florens Jelt., P. subpennatus Jelt., Isorthis perelegans (Hall), Gypidula ex gr. galeata Dalm., Leptostrophia rotunda Bubl., L. Sera Bubl., Leptaena ex gr. rhomboidalis W., Atrypa ex gr. reticularis L., Howellella mercuri (Gos.) kasachstanica Kap. Subsp. nom. mns., Bronteus sp. and plant remains of Psilophytaceae, 70 m.

6. Gray-green and brown-gray medium- and coarse-grained sandstones with thin beds of small-pebble conglomerates. Extrusive rocks predominate among the pebbles, 100 m.

7. Gray, tobacco-gray very fine-grained sandstones and sandy tuffs with Pholidostrophia sp., Anoplotheca sp., and Phacops ex. gr. Sternbergi C, 115 m.

[1]Section compiled by L.I. Kaplun and T.B. Rukavishnikova.

Plate I
Ludlovian brachiopods

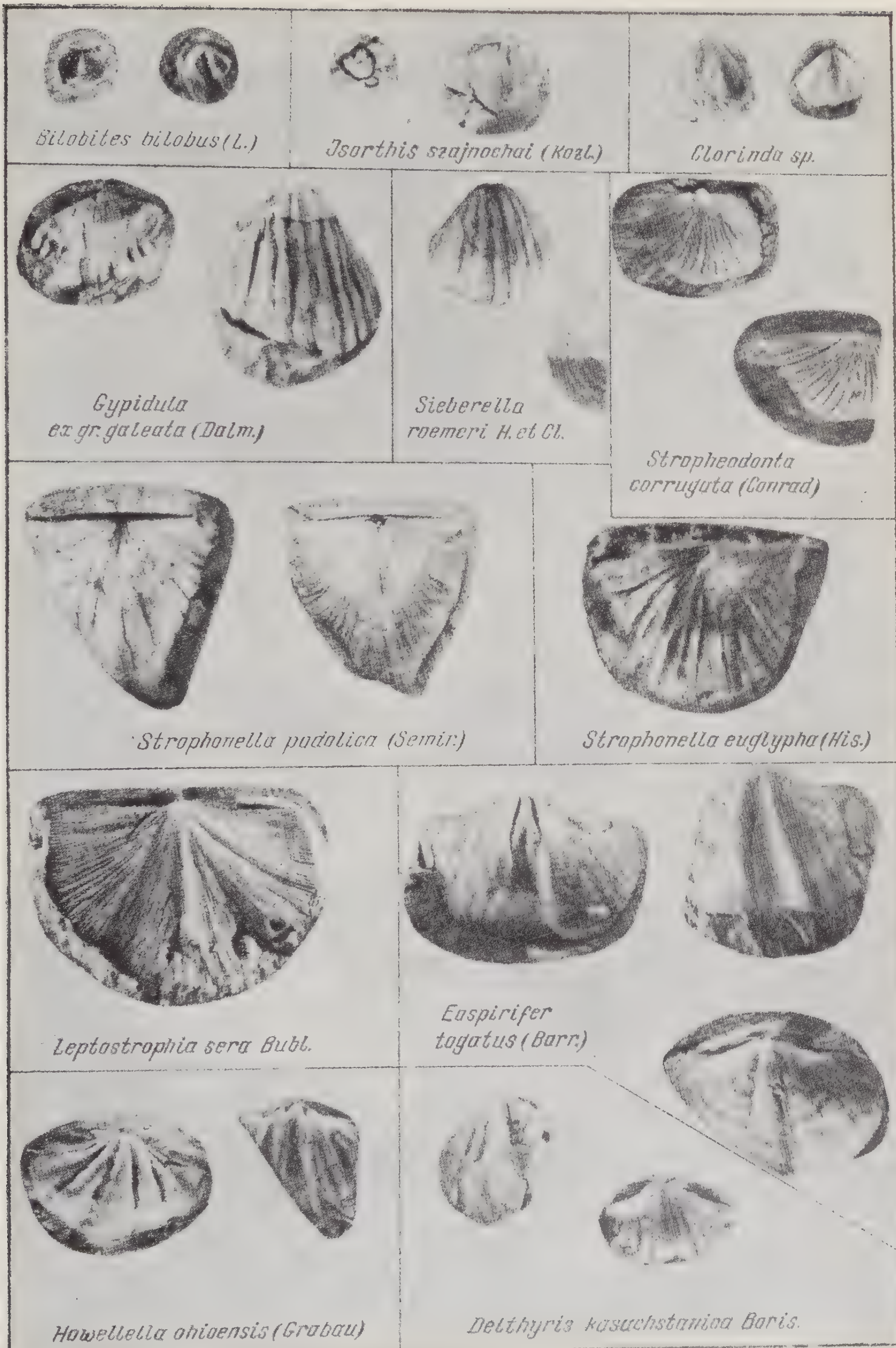


Plate II
Gedinnian Brachiopods

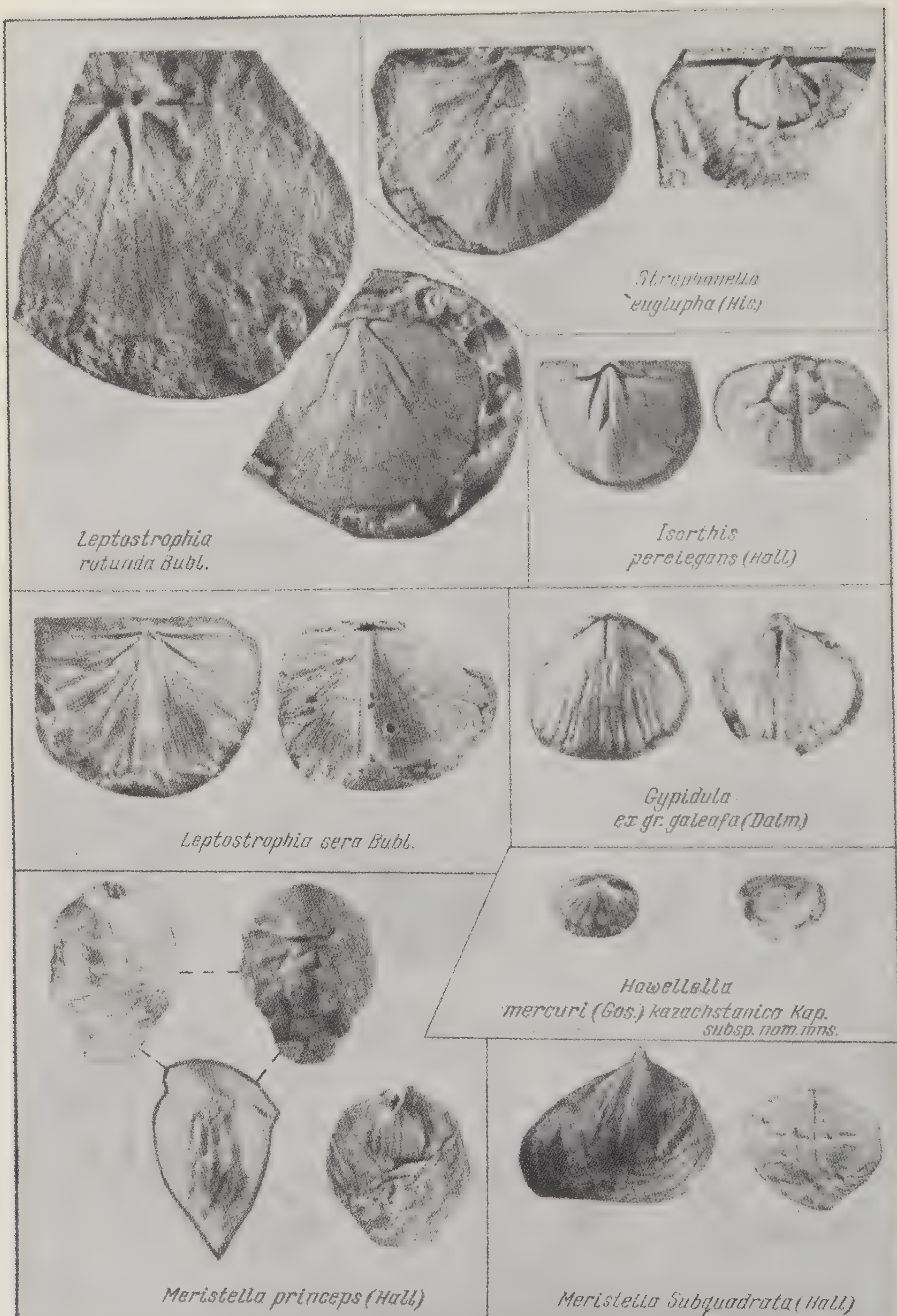
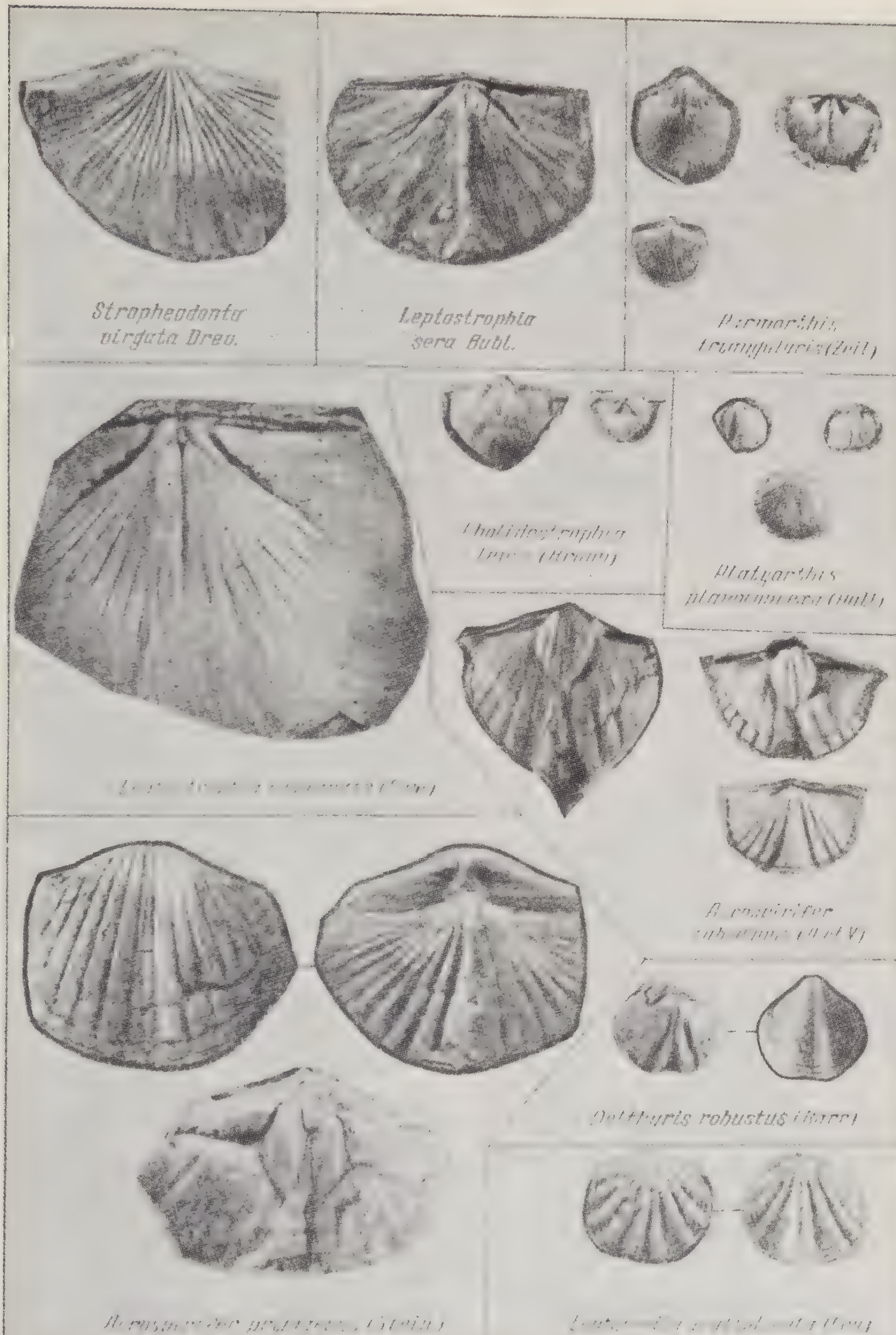


Plate III
Coblenzian brachiopods



8. Gray-green coarse-grained sandstones and sandy tuffs with thin layers of light-yellow tuffs, 120 m.

D_1^2 9. Green and yellow-green medium-grained sandstones and sandy tuffs with thin beds of conglomerate (20-50 cm) with andesite pebbles. The fauna of the sandy tuffs contains Chonetes sp. and Anoplotheca sp., 55 m.

10. Yellow-green fine-grained, less commonly coarse-grained sandy tuffs and sandstones with layers of tuffaceous sandstones. The sandstones contain abundant Pleurodictyum sp., Isorthis cf. perelegans (Hall), Leptostrophia magnifica Hall, Stropheodonta virgata Drev., Anoplotheca (?) sp., 110 m.

11. Gray and yellow-gray medium- and fine-grained calcareous sandy tuffs with beds of light-yellow tuffaceous sandstones. The calcareous sandstones contain abundant Parmorthis triangularis (Zeil.), Isorthis sp., Stropheodonta virgata Drev., Leptostrophia sowerby (Barr.), Schuchertella sp., Delthyris tetraplicatus var. grandis Kap. nom. mns., 160 m.

12. Tobacco-gray and gray very fine-grained sandstones, locally calcareous, interbedded with light-green tuffaceous sandstones. In the calcareous sandstones there are numerous Acrospirifer cabedanus (A. et V.) and Delthyris tetraplicatus var. grandis Kap. nom. mns., 150 m.

As can be seen from these sections, the Silurian and Lower Devonian sediments have very similar lithology, and key beds are absent. But the abundant faunas are found at all localities so that a boundary between Silurian and Devonian beds can be drawn on the basis of the change from one faunal assemblage to another.

An analysis of the paleontological material shows that there are three characteristic faunal assemblages[1] distributed in stratigraphic sequence and recurring with some variation in the species content in more than 20 separate sections recorded from different parts of the northeastern Balkhash region (Pl. L-43-8, 9, 20, 21, 22, 32, 33, 34, 35, 36).

Below are the lists of fossils in these assemblages (Tables I, II, and III).

The first assemblage: Syndetocrinus sp., Decacrinus pennatus Jelt., Chonophyllum sp., Pholidophyllum sp., Entelophyllum sp., Kyphophyllum ex gr. lindstromi Wdkd., Favosites (?) borissiakae Tschern., F. maubasensis Kov., F. Kelleri Kov., Heliolites decipiens M'Coy, H. subdecipiens Kov., H. stellaris Kov., H. Kuznetzkiensis (Tschern.), Helioplasmolites balticus Kov., Isorthis szajnochai Kozl., I ex gr. perelegans (Hall), Strophonella podolica (Semi.), S. Englypha (His.), Stropheodonta corrugata (Con.), Leptostrophia sera Bubl., Dictyonella sp., Bilobites bilobus L., Clorinda sp., Gypidula ex gr. galeata (Dalm.), Sieberella roemeri H. et Cl., Delthyris cf. kasachstanica Bor., Eospirifer togatus (Barr.), Atrypa ex gr. reticularis L., Howellella ohioensis (Grabau), H. ohioensis (Grab.) var. transversalis Ruk. var. nom. mns., Nucleospira sp., Dalmanites elegans Max. nom. mns., D. sp. a Max. sp. nov., Cheirurus quenstedti Barr., Phacops cf. rubidus Wdkd., P. aff. boecki Barr., Acidaspis sp., Lobopyge sp., Zosterophyllum sp., Taeniocrada (?) sp., abundant plant fossils of the Psilophytaceae family and imprints of primitive Lepidophyta.

The second assemblage differs considerably in its species composition from the first and contains: Decacrinus equilobatus Jelt., D. pennatus Jelt., Cyclocyclicus discoideus Jelt., C. spectabilis Jelt., C. gradatus Jelt., Pentagonopentagonalis subpennatus Jelt., P. florens Jelt., Isorthis perelegans (Hall), Strophonella euglypha (His.), Parmorthis balaensis Kap. nom. mns., Leptostrophia rotunda Bubl., L. sera Bubl., Gypidula ex gr. galeata (Dalm.), Howellella mercuri (Gos.), kasachstanica Kap. subsp. nom. mns., Delthyris tetraplicatus Kap. nom. mns., Meristella princeps (Hall), M. subquadrata (Hall), Nucleospira maillieuxi Dalm., Anoplotheca (?) sp., Phacops ex gr. logani Hall, P. ex. gr. sternbergi Corda, Dalmanites latepyge Max. nom. mns., Zosterophyllum sp.

The third assemblage is represented by: Cyclocyclicus echinatus Jelt., Pentagonopentagonalis florens var. magda Jelt., P. monstruosus Jelt., Pentagonocyclicus incusus var. kasachstanica Jelt., Kusbassocrinus kaplunae Jelt., Leptostrophia magnifica Hall, L. explanata (Sow.), Leptocoelia acutiplicata (Con.), stropheodonta virgata Drev., S. tenuivirgata Bubl., S. ampliata Bubl., Pholidostrophia lepis (Bronn.), Chonetes grandis Bubl., Parmorthis triangularis (Zeil.), Eospirifer sf. solitarius (Kr.), E. ignoratus Kap. nom. mns., Delthyris tetraplicatus var. grandis Kap. nom. mns. D. nimius Kap. nom. mns., D. robustus (Barr.), Acrospirifer primaevus (Stein.), S. assimilis (Fuchs), A. cabedanus (A. et V.), A. mediobalchaschensis Bubl., Nucleospira maillieuxi Dalm., Odontochile maccoy Barr., O. aff. ulrichi Delo, O. aff. fletcheri Barr., Crotalocephalus gibbus Beyr., Drepanophycus spinaeformis G., Protolepidodendron sp.

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[1] The main species of brachiopods in these assemblages are shown in Plates I, II, and III.

Table 1. Distribution of Some Silurian and Devonian Brachiopods in the Northeastern Balkhash Region

Epoch	Lower Devonian		Upper Silurian	Lower Silurian	
Stage	Coblenzian	Gedinnian	Ludlovian	Wenlockian	Llandoverian
Bilobites bilobus L			—— · · ·	....	
Isorthis perelegans (Hall)	——	——	——		
Isorthis sza nochai Kozl.		——	——		
Parmorthis balaensis Kap. non. mns.		——			
Platyorthis planoconvexa (Hall)	—— · · ·	—— · · ·			
Gypidula ex gr. galeata Dalm.		—— · · ·	—— · · ·	....	
Sieberella roemeri H. et Cl.		——	—— · · ·	· · ·	
Stropheodonta virgata Drev.	——				
Leptostrophia sera Bubl.	——	——	——		
Leptostrophia rotunda Bubl.		——			
Leptostrophia magnifica Hall	—— · · ·				
Leptostrophia beckii Hall	—— · · ·				
Strophonella podolica (Sem.)			——		
Strophonella euglypha (His.)		——	——		
Leptocoelia acutiplicata (Con.)	—— · · ·				
Eospirifer togatus (Barr.)			——		
Eospirifer solitarius (Kr.)	——				
Eospirifer ignoratus Kap. nom. mns.	——				
Delthyris kazachstanica Bor.			——		
Delthyris tetraplicatus Kap. nom. mns.	——	——			
Delthyris tetraplicatus var. grandis Kap. nom. mns.	——				
Delthyris robustus (Barr.)	——				
Delthyris nimius Kap. nom. mns.	——				
Hovellella mercuri kazachstanica Kap. subsp. hom. mns.		——			
Howellella ohioensis Grab.			—— · · ·		
Acrospirifer primaevus (Stein.)	——				
Meristella princeps (Hall)		—— · · ·			
Meristella subquadrata (Hall)		—— · · ·			

—— Distribution of the species in the northeastern Balkhash Region; .... in Western Europe; · · · · · in North America.

The first assemblage is characteristic of the Ludlovian, the second, of the Gedinnian, and the third, of the Coblenzian stage.

Below is an analysis of the brachiopod fauna of the Silurian and Devonian assemblages.

Since the grachiopod fauna of our region has not been sufficiently studied -- detailed studies exist for only a few species and many still undetermined species are omitted from the lists -- the analysis given here must be regarded as preliminary although it is sufficient to determine the age of the sediments.

A specific feature of the sedimentary rocks of this region is the mode of preservation of the brachiopods which are found mostly as internal and external casts and very seldom as original shells. This makes it difficult to compare them with the specimens from other regions where casts are almost always absent.

The Silurian brachiopods are represented by ten families: Bilobitidae, Schizophoriidae, Camerellidae, Pentameridae, Strophomenidae, Eichwaldiidae, Rhynchonellidae, Atrypidae, Coelospiridae, Spiriferidae. The most abundant, both in the number of species and number of specimens, are the following genera: Isorthis, Gypidula, Leptaena, Leptostrophia, Strophonella, Atrypa, Eospirifer, Delthyris.

The first brachiopod assemblage which establishes the upper Ludlovian age of the sediments contains, together with local species (Leptostrophia sera, Howellella ohioensis var. transversalis), such species as Isorthis szajnochai Kozl., occurring in the Borshchov (S_2^{1d}) and more rarely in the Skal'skiy (S_2^{1d}) strata of Podoliya, Sieberella roemeri H. et Cl., found in the Upper Silurian deposits of North America in the Brownsport and Henryhouse formations of North America; Stropheodonta corrugata (Conrad), characteristic of the Upper Silurian of North America (Rochester formation) (Strophonella podolica (Semi.), characteristic of the Borshchov formation of Podoliya and the Aynasuy beds (S_2^{1d}) of Central Kazakhstan [3]; Eospirifer togatus (Barr.),

Table 2.
The Distribution of Trilobites in the Upper Silurian and Lower Devonian Strata of the Northeastern Balkhash Region

Species	Upper Silurian	Lower Devonian	
		Gedinnian	Coblenzian
Cheirurus quenstedti Barr.	+		
Cheirurus ex gr. sternbergi Boeck.	+		
Cheirurus gibbus Beir.			+
Dalmanites nominalis Max. nom. mns.	+		
Dalmanites sp. α Max. sp. nov.	+		
Dalmanites latepyga Max. nom. mns.		+	
Dalmanites elegans Max. nom. mns.	+		
Dalmanites triangularis Max. nom. mns.			+
Phacops aff. boecki Barr.	+		
Phacops cf. rubidus Wdkd.	+		
Phacops ex gr. logani Hall		+	
Phacops ex gr. sternbergi Corda		+	
Phacops aff. fletcheri Barr.		+	
Phacops cf. fecundus var. communis Barr.	+		
Calcmene blumenbachi Brongn.	+		
Odontohile maccoy Barr.			+
Odontohile aff. ulrichi Delo			+
Odontohile aff. fletcheri Barr.		+	+
Scutellum aff. viator Barr.	+		

Table 3.
Distribution of Crinoids in the Upper Silurian and Lower Devonian of the Northeastern Balkhash Region

Species	Upper Silurian	Lower Devonian	
		Gedinnian	Coblenzian
Syndetocrinus sp.	+		
Scyphocrinus sp.	+		
Decacrinus pennatus Jelt.	+	+	
Decacrinus equilobatus Jelt.		+	
Cyclocyclicus discoideus Jelt.		+	
Cyclocyclicus gradatus Jelt.		+	
Cyclocyclicus spectabilis Jelt.		+	
Cyclocyclicus echinatus Jelt.			+
Pentagonopentagonalis florens Jelt.		+	
Pentagonopentagonalis florens var. magna Jelt.			+
Pentagonopentagonalis subpennatus Jelt.		+	
Pentagonopentagonalis monstruosus Jelt.			+
Pentagonocyclicus radialis Jelt.		+	
Pentagonocyclicus incisus var. kazachstanica Jelt.			+
Pentagonocyclicys singularis Jelt.		+	
Kusbassocrinus kaplunae Jelt.			+

known from the Wenlockian to the Upper Ludlovian[1] (ed-ej) of Czechoslovakia; Delthyris kasachstanica, which occurs in the Aynasuy beds of Central Kasakhstan; and Howellella ohioensis (Crab.), common in the Upper Silurian of North America (Monroe formation).

Besides the Upper Silurian species there are species with a greater vertical range such as Bilobites bilobus (L), known throughout the Silurian; Gypidula galeata (Dalm.) and Strophonella euglypha (His.), formerly known from Silurian strata only but occurring in the northeastern Balkhash region in the Gedinnian stage as well.

In the sediments immediately above the horizon with the brachiopod fauna the following corals were found (Pl. L-43-8, Nurashbasy, Pl. L-43-21, Kokbaytal, Pl. L-43-33, Maubas): Favosites (?) borissiakae (Tschern.), F. maubasensis Kov., Heliolites decipiens M'Coy, H. subdecipiens Kov., H. stellaris Kov., H. Kuznetskeinsis (Tschern.), and Helioplasmolites balticus Kov. According to O. P. Kovalevskiy [5], this assemblage of tabulates and Heliolites indicates that the sediments are of upper Ludlovian age. Thus, the conclusion that the age of the sediments is upper Ludlovian based on the brachiopod fauna is confirmed by the coral fauna. The Lower Devonian brachiopods in these beds are represented mainly by the Spiriferidae and Strophomenidae and to a less extent by Watsellidae, Schizophoridae, Chonetidae, Atrypidae, Coelospiridae, and Meristellidae.

The second brachiopod assemblage which served as the basis for separating the Gedinnian stage in the northeastern Balkhash region includes, besides the local species (Leptostrophia rotunda Bubl., Parmorthis balaensis Kap. nom. mns., Delthyris tetraplicatus Kap. nom. mns.), such species as Isorthis pereleganс (Hall), Meristella princeps (Hall), M. subquadrata (Hall), which are common in the lower Helderberg beds (D_1^1) of North America, and Howellella mercuri (Gos.) kasachstanica Kap. subsp. nom. mns., which is a geographical subspecies of Howellella mercuri (Gos.), characteristic of the Gedinnian stage.

[1]According to an oral communication from the Czech paleontologist R. Garny, Eospirifer togatus (Barr.) is found in Czechoslovakia from the Wenlockian to the Upper Ludlovian (ed-ej) and does not continue into Devonian as was once believed.

of the Ardennes (Mondrepuys shales).

The second brachiopod assemblage occupies an intermediate position in the section and contains, besides the species characterizing the lower Helderbergian beds of North America and the Gedinnian stage of western Europe, some elements of the underlying (Ludlovian) and the overlying (Coblenzian) faunas.

The Ludlovian species present in it are Leptostrophia sera Bubl., Strophonella euglypha (His.), and Gypidula ex. gr. galeata (Dalm.). Leptostrophia sera Bubl. extends higher and is found in the typical Coblenzian faunas.

The presence of the Gedinnian forms, the mixed character of the second brachiopod assemblage, its position in the section, and structural and lithological similarities permit us to separate the Gedinnian stage in the section, and to speak of a gradual change from Silurian to the Devonian.

The third brachiopod assemblage is characterized by an exceptional abundance and variety of the brachiopods which differ sharply, generically and specifically, from those of the underlying assemblage and by the great development of the representatives of the spiriferids, especially of the Acrospirifer, Eospirifer, and Delthyris genera.

Besides the many endemic forms, such as Stropheodonta ampliata Bubl., S. tenuivirgata Bubl., Chonetes grandis Bubl., and others, there are brachiopods, characteristic of the Coblenzian stage of western Europe, such as Leptostrophia explanata (Sow.), Eospirifer sf, solitarius (Kr.), Acrospirifer primaevus (Stein.), A. assimilis (Fuchs.), and of the Oriskany formation (D_1^2) of North America, such as Leptostrophia magnifica Hall and L. beckii Hall.

The gradual changes in the fossil assemblages indicate that at the boundary between the Silurian and Devonian, there existed in our region a marine basin in which lithologically uniform tuffaceous-sedimentary materials were deposited without interruption.

However, interruptions in sedimentation must have occurred locally, for some of the Devonian stages are absent from individual sections; for example, in the Shaintas Mountains (Pl. L-43-21) the Gedinnian stage is absent and the Coblenzian sediments rest, with a basal conglomerate, on the Ludlovian sediments. Such erosional disconformities and interruptions of sedimentation were the result of, apparently, differential oscillatory movements continuing in our region through the entire Devonian period.

SUMMARY

1. The Silurian and Devonian sediments of the northeastern Balkhash region are conformable and enter into the same geological structures, indicating that no orogenic movements occurred at the Siluro-Devonian boundary.

2. In the Silurian-Lower Devonian time, the region was occupied by a marine geosynclinal basin. Frequent oscillations of the basin bottom changed the configuration of the islands and gulfs and the distribution of land and sea areas. The variation in the thickness of Silurian and Devonian sediments and local disconformities are the result of these movements.

3. The Silurian and Devonian terrigenous deposits are tuffaceous clastics, indicating that the marine basin was shallow. The abundance of fossils (crinoids, corals, brachiopods, trilobites, pelecypods, and gastropods) indicates that the sea had normal salinity and temperature. The numerous plant remains indicate that the sea was epicontinental.

4. The absence of extrusive igneous rocks and crystal tuffs suggests that volcanoes were no longer active in the region during the Silurian-Lower Devonian time; the abundant volcanic ash was probably transported from the western Balkhash region and from central Kazakhstan.

5. The abundance of fossils makes it possible to establish the presence of fossil assemblages characteristic of the upper Ludlovian, Gedinnian, and Coblenzian ages. These three assemblages of different ages pass gradually one into another and therefore the Siluro-Devonian boundary can be drawn only on the basis of the change in these assemblages.

6. The upper Ludlovian brachiopod assemblage of the northeastern Balkhash region resembles the fauna from the Ainasui beds of central Kazakhstan.

7. The Silurian and Lower Devonian faunas of the region, contain North American and western European elements and a large number of forms found only in the local zoogeographic province. The forms common to both western Europe and North America have a somewhat different vertical distribution in our region and are differently associated in the assemblages, but with a slight generalization the sections of the northeastern Balkhash region can be correlated with the western European subdivisions of the Silurian and Devonian periods.

Bublichenko [2] has suggested a local

cheme of subdivision of the Lower Devonian or our region into Pribalkhash beds tentatively orrelated with the Gedinnian stage, and the ardzhal'skiy beds correlated with the Coblenian stage of western Europe.

A local stratigraphic column is very much eeded because the northeastern Balkhash egion is not entirely analogous to western urope or North America either in the fossil ssemblages or in the vertical distribution of he species. But since the Balkhash fauna is till insufficiently known, and not all of the roups have been studied in detail, we conider it advisable to retain the accepted estern European stratigraphic subdivisions f the Silurian and Lower Devonian, at least or the present.

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THE IMPORTANCE OF QUANTITATIVE DETERMINATION OF COLOR IN THE STUDY OF SEDIMENTARY URANIUM DEPOSITS

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The paper presents data on the exact photometric determination of color in uranium-bearing carbonate rocks and analyzes the relation between color and the content of uranium compounds, organic matter, and various forms of iron in the rocks.

The author concludes that the color of sedimentary rocks is an important indicator of the geochemical environment of their formation and therefore of the presence or absence of conditions favoring fixation of uranium compounds.

* * * * *

Many methods are used in modern geology in studying the composition of sedimentary rocks and associated ore deposits. One of the most important characteristics of sedimentary rocks is their color. Being intimately related to definite constituents of rocks, it commonly simplifies their identification and in many cases characterizes the medium in which a given rock was formed. It is well-known, for example, that the greenish-gray hues are the result of the presence of ferrous compounds in minerals of the glauconite type and indicate that the rock was formed in a slightly reducing environment; brown color is due to hydrous iron oxides and indicates oxidizing conditions, etc.

Inasmuch as sedimentary ore minerals are often syngenetic and many of them are very sensitive to changes in environment, it is evident that color may be as valuable a guide to ore as the oxidation potential, hydrogen ion concentration, and other geochemical guides. However, in using color for this purpose, approximate visual determination such as is usually made in describing sedimentary rocks is quite inadequate.

The need for a quantitative determination of color has been discussed in recent years by a number of investigators. Various methods of studying color have been described both in the Soviet and foreign literature [1, 4, 5, 8, 16, 17, 21]. Actual application of these methods shows that the most convenient of these methods is the optical photometric method described in a number of papers and in a recently published manual of sedimentary petrography [5, 15].

The photometric method gives the color of any specimen or a channel sample, in terms of a series of numbers, just as do mineralogical, chemical, and other analyses of rocks. This makes it possible to correlate the color of a rock with its composition and construct chromographs, color maps, and sections of individual stratigraphic beds. In spite of the obvious advantages of photometric measurements, they have not yet found wide recognition in geologic practice.

Below we present a number of examples of the use of color in the study of various types of uranium deposits in carbonate and terrigenous sediments. The results of measurement of samples of uranium-bearing carbonate rocks with the universal photometer FM were used for correlating the color of rocks with their composition, and for making a combined chromatic and radiometric survey. The results of this survey and the analysis of data obtained from it are shown in Fig. 1. The upper diagram presents a number of sections along the exposed part of the ore-bearing bed. The carbonate rocks which contain uranium ore at this locality, usually in the form of "thucholite" and pitchblende, are mainly oolitic and fine-grained biogenic limestones. The limestones are characterized by the presence of stylolites and the greatest accumulations of the achromatic color characteristic. The correspondence between these

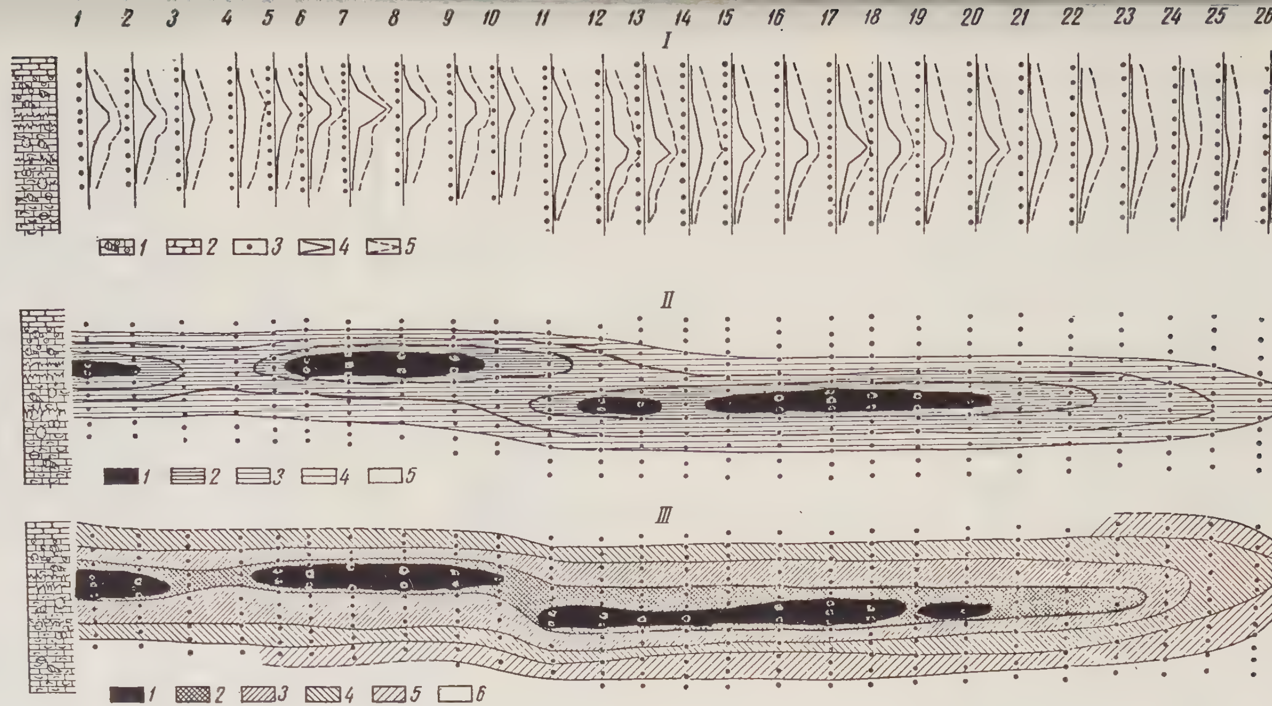

FIGURE 1. Bedded-lenticular ore bodies in sedimentary uranium deposits in carbonate rocks (prospect cuts 1 - 26)

I -- prospect cuts in the ore-bearing horizon: 1 -- limestones; 2 -- dolomites, 3 -- place of sampling; 4 -- curve of variation in uranium content for each section; 5 -- curve of variation in reflectivity of samples in each section.

II -- ore bed with curves of equal uranium content: 1 -- very rich ore; 2 -- rich ore; 3 -- ore with average uranium content; 4 -- poor ore; 5 -- rocks with average uranium content.

III -- ore bed with lines of equal reflectivity of rocks. Percentage of black (S): 1 -- over 50%; 2 -- 50 to 40%, 3 -- 40 to 30%; 4 -- 30 to 20%; 5 -- 20 to 10%; 6 -- less than 10%.

two graphs is evident; the areas of the bed containing a higher percentage of uranium are, as a rule, distinguished by the darker color of the enclosing rocks.

In order to show this relationship more clearly, the result of photometric measurements (percentage of black) and the uranium content in all 308 samples were plotted on a graph in rectangular coordinates (Fig. 2). Here the scatter of points is considerable but the general regularity of the color change in rocks with the increase in uranium content is sufficiently clear, especially if the average content of uranium for each group of samples of similar color is computed (10 to 20 percent black, 20 to 30 percent black, 30 to 40 percent black, etc.). The curve illustrating this relationship is also given in Fig. 2.

sence of uranium minerals, which are mainly black. However, the very small content of these minerals suggests that they are not the principal chromophores. It is known that the variation in color of the carbonate rocks may be also the result of the presence of clastic material. In our rocks, the content of clastic material ranges from tenths of one percent to a few percent in the purer varieties and as much as 20 percent in the varieties most strongly enriched in terrigenous material. Most commonly the detrital material consists of such stable minerals as quartz and the less abundant feldspars and mica. Because of their light color, these minerals cannot noticeably affect the color of the rock.

The content of the dark minerals, hornblende, pyroxenes, and the accessory min-

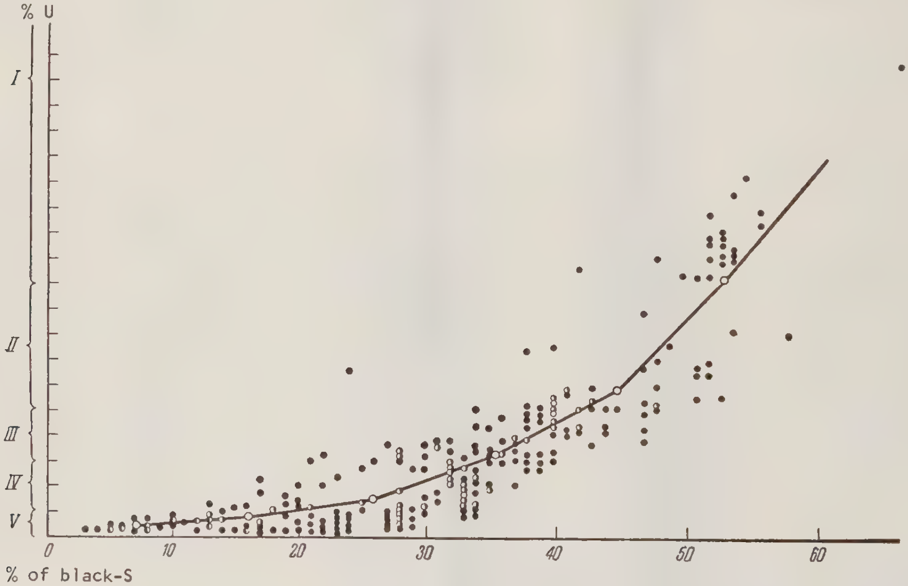

FIGURE 2. Color (achromatic, % of black -- S) of the carbonate rocks with various uranium content: I -- Very rich ores; II -- Rich ores; III -- Ores with an average uranium content; IV -- low grade ores; V -- rocks with uranium content near the clarke value.

To determine the source of color of a sedimentary rock, it is evidently necessary to determine the color effect of every one of its constituents.

The natural or, to use A. Ye. Fersman's term [13], "idiochromatic" color of limestones without any admixtures is probably white. Any other color in a limestone is the result of various admixtures, sometimes present in very small amounts. The gray color in the limestones under discussion may evidently be ascribed to the pre-

erals such as zircon, garnet, tourmaline, staurolite, rutile, and others is usually negligible and they, too, cannot have any significant effect on the color changes in the rock mass. In searching for other chromophores capable of coloring the uranium-bearing rocks, attention was given to the organic content of the rocks. Organic matter is one of the common constituents of sedimentary rocks and has a very important effect on the geochemical processes occurring in them [2]. The organic content of the sedimentary rocks has been investi-

gated in detail a number of times. The best known investigations are those by P. Trask and H. Patnode [21, 22], who studied the petroliferous rocks from a number of provinces in the U.S.A.

The results of their investigations are shown in graphs and diagrams illustrating the relation between color of the rocks and the content of organic matter in them. The amount of the latter was determined either by direct analysis for organic carbon in the samples [22 and Fig. 3] or by nitrogen determination and computation of the so-called reduction number [21, Fig. 4]. The samples were first divided by the authors into 37 groups according to the intensity of the dark color. The content of organic carbon in percent was plotted on the ordinates (Fig. 3) and the relative frequency of the samples characterized by a given carbon content in each of the 37 groups was plotted on the abscissas. The means for each of the individual graphs are shown by semicircles and the curve drawn through the graphs shows the increase in the intensity of the dark color with increase in the content of organic carbon. In determining the organic matter in core samples, Patnode [21] oxidized it with an 0.4 N solution of chromium anhydrite and called the amount of this substance needed to oxidize 100 mg of the sample the reduction number. The results of determination of the reduction number and of the reflectivity of the core samples from one of the drill holes are given in Fig. 4. It clearly shows the correspondence between the two indicators.

The change in the color of the carbonate rocks with their organic content was studied in the ore-bearing horizons shown in Fig. 1. The relation between the color of the rock and its organic carbon content is shown in Fig. 4 and is based on the study of 111 samples of uranium-bearing and barren limestones and dolomites. The amount of organic carbon in these rocks ranges from 0 to 0.63 percent, and the percentage of black color, from 10 to 70 percent. Thus, the color of the rocks ranges from light-gray, almost white, to gray and dark-gray. A considerable scatter of points evidently indicates that in this case the organic matter is not the only chromophore determining the color of the rock, although its role in modifying the idiochromatic color of the investigated series of carbonate rocks is quite important. The curve connecting the mean values in Fig. 5 shows clearly the general relation between the darkening of the rocks and the increase in organic carbon content.

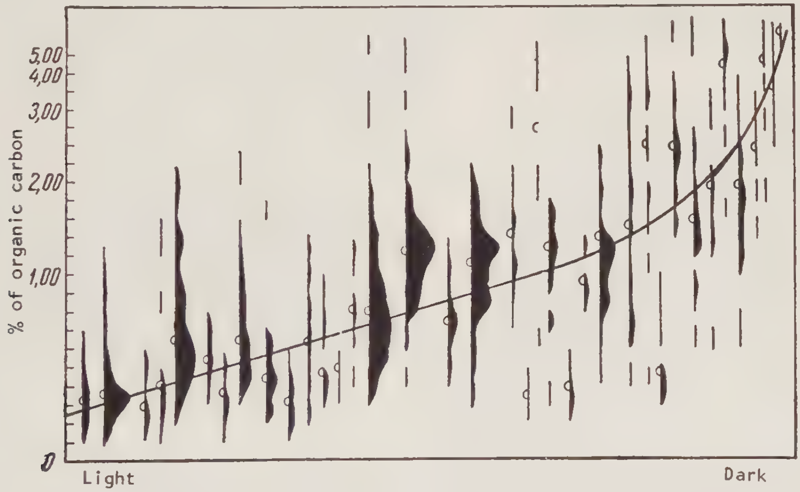

FIGURE 3. Relation between color and the content of organic carbon in sedimentary rocks (after P. Trask and H. Patnode, 1936)

When the data on the organic carbon content are compared with the data on the uranium content in the samples, a definite relationship is revealed. Two groups of rocks can be distinguished in Fig. 6. In one group the increase in uranium content corresponds to the increase in organic carbon content; in the other group, which falls on the abscissa of the graph, the increase in uranium content does not accompany the

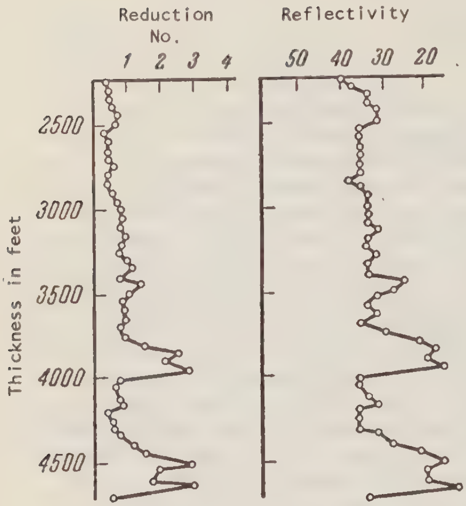

FIGURE 4. Vertical variation in the reduction number and reflectivity of samples (after H. Patnode, 1941)

increase in organic carbon content. Luminescent analysis for bitumens and chemical analysis of organic matter in these two groups of rocks showed that the nature of organic matter is substantially different in the two groups. When the content of organic carbon exceeds 0.4 to 0.9 percent, the rocks usually contain bitumens in thin fractures, pores, and cavities. A smaller content of organic carbon is due, apparently, to finely dispersed organic matter which may accumulate in sedimentary rocks during deposition. Thus, a connection is suggested between the concentration of uranium into ore and the dispersed, evidently primary, organic matter.

Considering other possible causes of coloration of limestones, it is necessary to analyze the state of iron in its various compounds which, as is well known, are among the most important coloring agents of sedimentary rocks. Their importance in coloring clays was first demonstrated by C. Tomlinson [23]. A similar work on limestones was done by the present author. The samples in which the content of ferrous and ferric iron was determined were taken from two stratigraphic horizons. One of these contains uranium ore, mainly in the form of pitchblende and thucholite; the other has no uranium mineralization at all. In all samples the total content of ferrous and ferric iron is seldom as much as 0.9 percent. In most of the samples (95 out of 113), the amount of ferrous and ferric iron does not exceed 0.5 percent. The samples from the uranium-bearing carbonate rocks are mainly different shades of gray; the samples from the barren rocks are light-yellow, light pink, and yellowish-pink.

The substantial difference in color of the ore-bearing and barren samples, because of the difference in the ratio of the ferrous to the ferric iron, is clearly shown in the distribution of these analyzed samples on the diagram (Fig. 7), which is constructed like Tomlinson's diagram [23]. In the

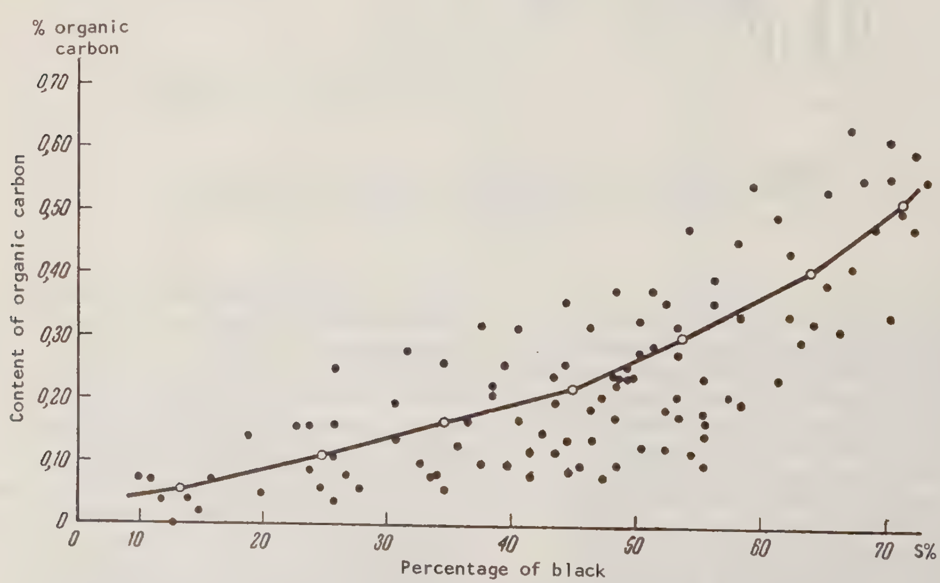

FIGURE 5. Change in the color of carbonate rocks (limestones and dolomites) in relation to the content of organic carbon (based on analyses of 111 samples)

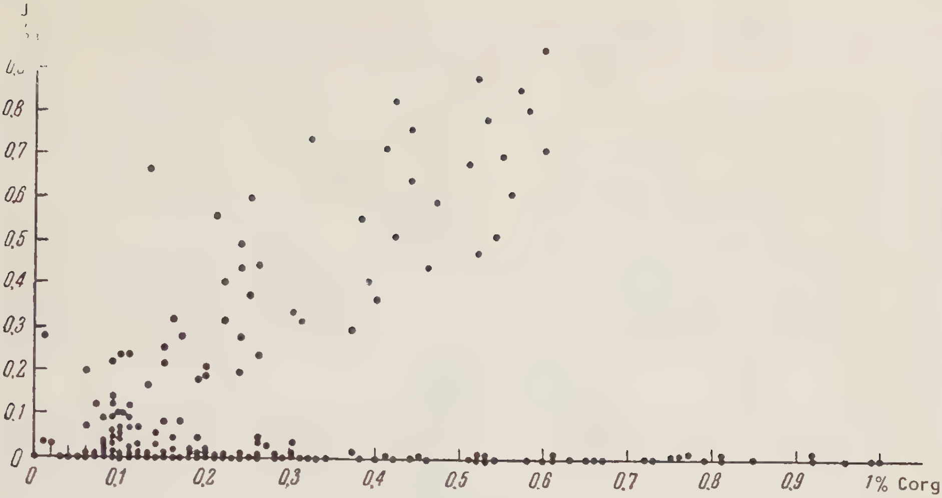

FIGURE 6. Relation between Uranium content (U) and organic carbon content (Corg) in carbonate rocks of the ore-bearing horizon.

Based on analyses of 180 samples containing as much as 1% of each component. Samples with higher content of the two components were excluded as having obviously been subjected to secondary enrichment.

samples distributed along the vertical axis of the diagram, from base upwards, i.e. in the order of decrease of Fe_2O_3 from 0.70 to 0.001 percent, there is a gradual change in color from pink and yellowish pink to light-yellow; in the samples clustered about the horizontal axis, the ferrous iron predominates and varies from 0.02 to 0.92 percent; here also the color changes gradually but in achromatic hues from light-gray to dark-gray (from left to right).

The diagram shows also that the ore-bearing rocks are sharply separated from the barren and cluster about the horizontal axis, thus emphasizing the predominance in them of ferrous over ferric iron. This, in its turn, indicates an intimate connection between uranium mineralization and the reducing environment. The diagram contains even more profound genetic implications. The sharp separation of the two kinds of samples into two independent branches following the vertical and the horizontal axes of the graph points to the fact that the chemical differentiation of matter in the sediments [9, 12] was not limited to the time of deposition but continued into the diagenetic stage, which consists in reactions between the sediments and the surrounding medium [12]. These reactions and the following compaction and cementation, i.e., lithification of sediments, led in one case to the dominance of reducing conditions and the change of ferric iron compounds into (with the formation of sulfides, $FeS_2 . \underline{n}\ H_2O$ and finally of $Fe_2O_3 . \underline{n}\ H_2O$).

The reducing environment favored the accumulation and preservation of uranium compounds in the form of black uranium oxides, pitchblende, etc.; the oxidizing environment encouraged their migration. An examination of the distribution of points representing ore-bearing samples on the graph brings out another interesting feature. In most of the samples ferrous iron predominates strongly over ferric iron; however, in some samples (in 7 out of 85) the opposite relation is observed (dominance of ferric iron). On the graph these samples lie to the left of the diagonal line representing the 1 to 1 ratio of ferrous and ferric iron. A mineralogic analysis showed that the uranium in these samples is mainly in the form of uranium vanadates. These samples were taken from the oxidized parts of the bed in which [11] uranium vanadates (carnotite and tyuyamunite), shroekingerite, uranotil, and other minerals are the most stable.

These data indicate that the dark colors of uranium-bearing carbonate rocks are produced by a number of substances, the most important of which are organic matter and ferrous compounds. These substances are also indicators of the reducing geochemical environment in the rocks which enclose them. In most investigated samples there is a direct relationship between these

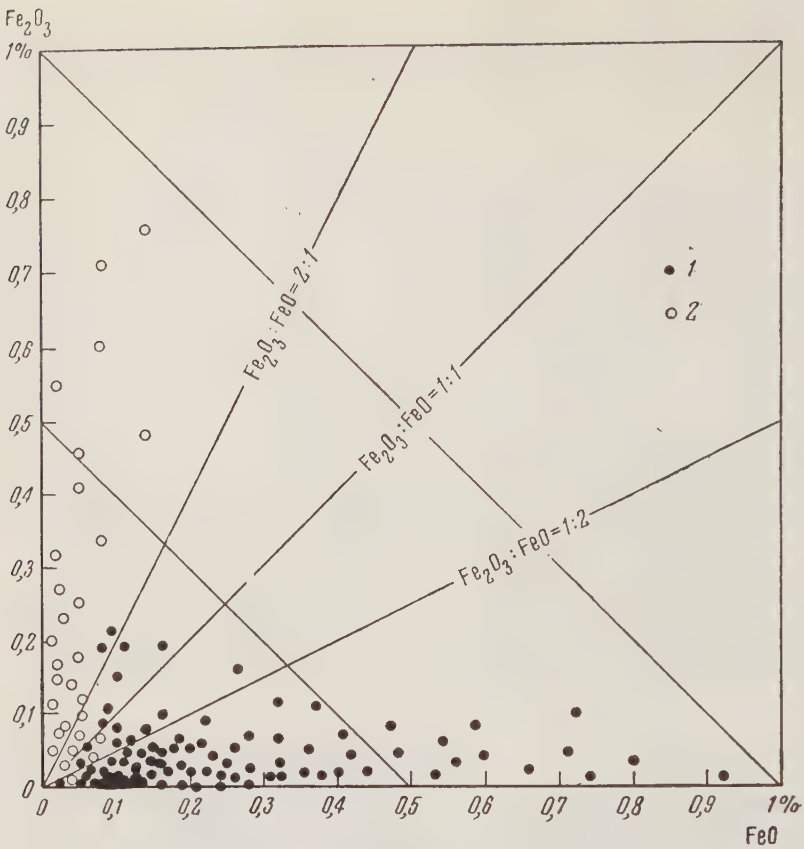

FIGURE 7. Relation between the content of ferrous and ferric iron in carbonate rocks (based on analyses of 113 samples).

1 -- Ore-bearing horizon (84 samples).

2 -- Barren horizon (29 samples).

compounds and uranium mineralization, indicating that reducing conditions favor concentration of uranium.

Interesting data showing the significance of color of sedimentary rocks as a guide to sedimentary uranium deposits are to be found in the paper by J.A. Masters [19]. The author gives a brief description of the ore-bearing flood-plain sediments, largely Upper Jurassic red sandstones, siltstones, and clays. Within them there are sandy lenses representing "paleostream channels." These channel deposits consist of gray and brown-gray sandstones which are coarser, less consolidated, and more permeable than the enclosing redbeds.

Uranium mineralization, the author points out, is localized in these gray and brown sandstones. He regards their gray and brown color as a secondary feature produced by the passage of mineralizing solutions through the more permeable parts of the sediments. From the data provided by drill cores, the author calculated the thickness of gray and brown beds "favorable" and the redbeds "unfavorable" to ore deposition. Referring these data to unit thickness, Masters drew color maps using isopercentages of "favorable" color for the contours. One of these maps is shown in Fig. 8.

The color maps drawn by Masters [19] for the deposits of his region show also the characteristics of the distribution of the gray paleostream deposits among the flood-plain redbeds. In the central part of Fig. 8, an area of channel deposits is outlined by the 100 contour (100 percent "favorable" color in terms of the chosen unit of thickness). The 75.50 and 25 contours indicate a gradual (sometimes rather abrupt where the contours are crowded, sometimes smooth where they are farther apart) wedg-

FIGURE 8. Color map with contours showing the content of sandstones with "favorable" color in one of the Arizona regions (after J. A. Masters, 1955)

1 -- Ore body; 2 -- Ore hole; 3 -- Mineralized hole; 4 -- Barren hole; 5 -- Formation boundary.

g out of gray sandstones and increasing ickness of the redbeds. The drill holes ntaining ore (economic mineralization), les with non-economic uranium mineralization, and barren holes are also shown in Fig. 8 by appropriate symbols. All ore-bearing and mineralized holes are located in the area of sandstones with "favorable"

color, while the extensive areas of redbeds are completely devoid of any signs of uranium mineralization.

The areas where the paleostream channel deposits are widespread (lower part of Fig. 8) contain separate ore bodies and numerous mineralized holes. Masters [19] considers that "persistent gray sandstones appear to be relatively unfavorable (for the formation of economic ore bodies, V. D.) because in them the mineral concentration tends to be dispersed." But where the ancient channels become narrow (middle of Fig. 8) and are gradually replaced by redbeds, ore bodies are usually of considerable size. The author draws attention to the definite orientation of the ore bodies either parallel to bedding or coinciding with the general trend of the ancient channels. Without discussing Masters' ideas on the genesis of the ore, it must be admitted that the localization of uranium mineralization in rocks with a definite "favorable" color is very convincingly presented in his work.

The value of color of sedimentary rocks as a guide to ore and the origin of color in uranium deposits is no less clearly presented in the paper by R. Vickers [24]. Discussing the Lower Cretaceous uraniferous sandstones, he draws attention to the connection between uranium mineralization and radioactive anomalies and the color of the enclosing rocks. This is illustrated by a number of maps and sections, one of which is reproduced in Fig. 9. The wavy line in the middle of the section indicated the red-buff contact within a 1 1/2 m sandstone bed. The lenses of carnotite-bearing sandstones are enclosed in light-gray sandstone and lie near the red-buff contact. (The colors are named by P. Vickers according to the Rock Color Chart, National Research Council, U.S.A.) The high concentrations of uranium minerals at the zone of color change is characteristic for the region described by Vickers (Fig. 10). The presence of this zone, Vickers points out, made it possible to locate a number of new uranium occurrences in outcrops and in mines.

As for the cause of color change in the sandstones and the formation of high concentrations of uranium minerals, Vickers holds the following view. He believes that primary dispersed uranium existed in the redbeds in association with carbonaceous material and pyrite. The presence of the latter in the altered sediments is proved by the presence of goethite pseudomorphs after pyrite. As a result of pre-Oligocene weathering, uranium compounds were changed into hexavalent form and migrated down dip. The total content of iron in sandstones of different colors ranges from 0.1 to 0.5 percent. The red color of the sandstones is explained as a result of oxidation of minerals containing ferrous iron and yellow-brown hydrous iron oxides into hematite.

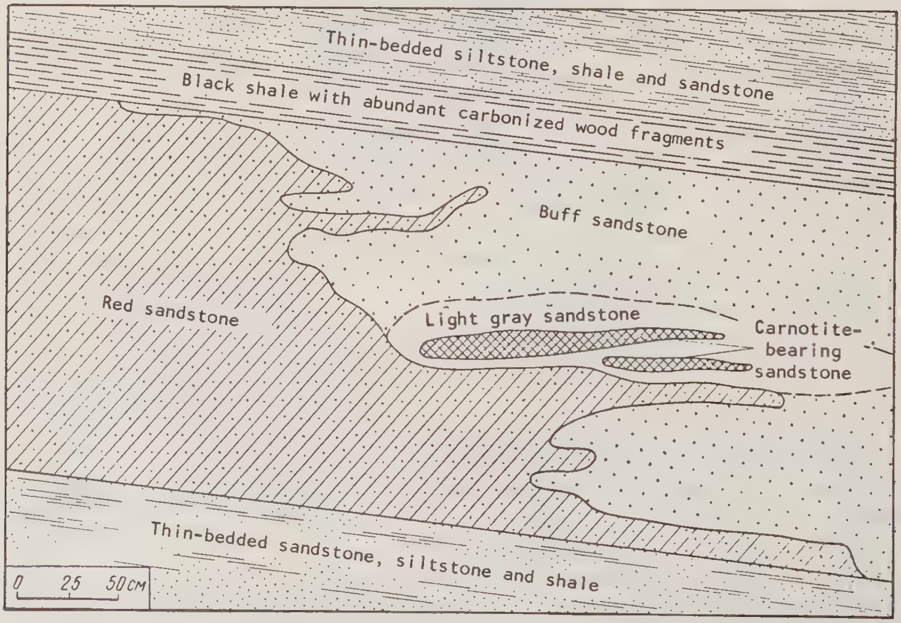

FIGURE 9. A section showing the relation between red-gray contact and carnotite-bearing sandstone (after R. Vickers, 1957)

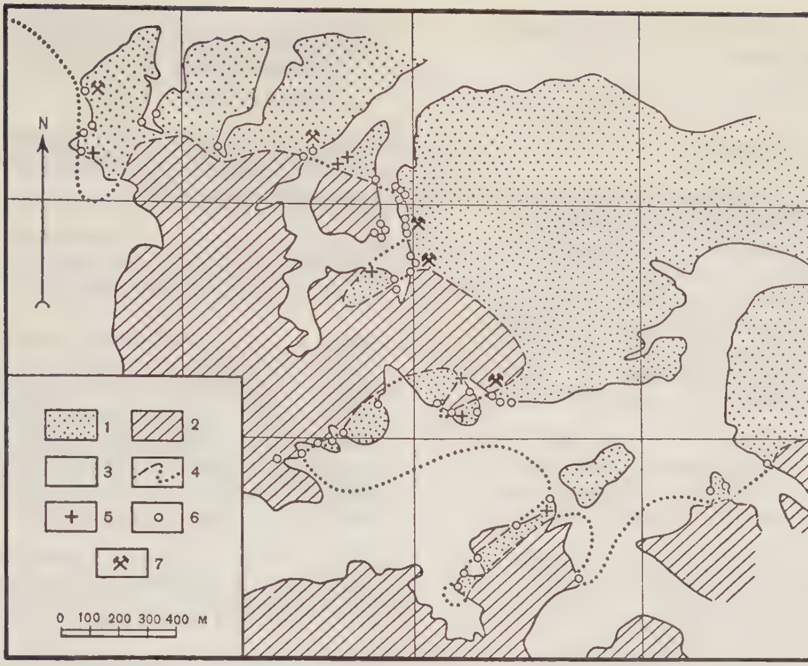

FIGURE 10. Map showing relationship of color change in sandstones, radioactivity anomalies, and uranium mineralization (after P. Vickers, 1957)

1 -- Buff basal sandstone; 2 -- Red basal sandstone; 3 -- Eroded basal sandstone; 4 -- Boundary of color change between buff and red sandstone; 5 -- Uranium mineralization; 6 -- Radioactivity anomalies; 7 -- Mine workings.

e boundary of the two geochemical envi-
onments.

The observations of Masters [19] and ickers [24] on color change in terrigenous ranium-bearing sediments is interesting oth in connection with the problem of the rigin of uranium mineralization and in con- ection with prospecting. This is clear om the color maps and sections drawn by ese authors, which show that uranium ineralization is localized in the zone of olor change (Figs. 8, 9, 10). We cannot, owever, approve the method used by P. icker in determining colors. Such a term s "buff sandstone," sandstone of the color f buffalo leather, does not clearly define e color of the sandstone; and the use of olor charts [24] as standards for naming olors is not general enough and makes it npossible to use the cited data for com- arison. This emphasizes again the need f determining colors by an optical photo- eter rather than visually.

Favorable conditions for the concentra- on of uranium which are reflected in the olor of the rocks may have occurred at fferent stages in the history of the rocks. sedimentary rocks containing primary organic material, uranium may be concentrated during the early stages of deposition and diagenesis. In carbonate rocks these early concentrations of uranium compounds may sometimes be of practical interest, although the formation of ore bodies is completed, as a rule, at a later stage; but economic concentrations of uranium in terrigenous rocks are usually related to later hydrochemical processes which destroy the poor primary uranium mineralization in one part of the bed and concentrate uranium in another. Especially characteristic of the terrigenous deposits is the formation of economic uranium concentrations at the boundary between two different geochemical environments in the ore-bearing rocks as described by Vickers [24].

These examples of the study of color of carbonate and terrigenous uranium-bearing rocks show that in a number of cases color or a sharp color change indicates a change in geochemical environment and may serve as a guide to ore. For this reason, precise determination of the color of rocks and construction of chromographs of sections and color maps for individual stratigraphic horizons must become one of the phases of complex geological investigations

undertaken in the search for uranium deposits in sedimentary rocks.

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ON THE POSITION OF THE RUDNYY ALTAI IN THE STRUCTURAL PLAN OF THE SAYAN-ALTAI REGION

by

A. L. Matveyevskaya

The paper compares the geologic history of the Kolyvan'-Tomsk and Irtysh-Zaysan (or Kalba) parts of the Ob-Zaysan Variscan geosynclinal system. On the basis of this comparison and analysis of data on the structure and geologic history of the Rudnyy Altai and neighboring regions, the author concludes that the Rudnyy Altai consists of small block structures developed at the juncture of the Variscan geosynclinal system and the epi-Caledonian platform of the Gornyy Altai, which borders it in the east.

* * * * *

On the basis of many years' work of V.P. Nekhoroshev [12, 13, 14], N.L. Bublichenko [3], N.A. Yeliseyev [8], and others, the Rudnyy Altai was considered as the "shelf zone" of the Zaysan (or, in Soviet terminology, Ob-Zaysan) Variscan geosyncline. It was believed that this shelf zone lay between the Caledonian folds of the Gornyy Altai, which was a platform in Variscan time, and the central part of the above-mentioned geosyncline, which existed at the site of modern Kalba. At the end of the Late Paleozoic this region was subjected to intensive folding in two stages accompanied by large intrusions. Later movements thrust the Gornyy Altai and the Kalba over the Rudnyy Altai along the dislocations which formed at that time in the Irtysh and the northeastern zones of deformation.

Approximately the same position was ascribed to this zone in the later works of G.D. Azhgirey and P.F. Ivankin [1], Yu.A. Kuznetsov [9], V.A. Kuznetsov [10, 11], and others, but these authors emphasized the great differences in the Middle and Upper Paleozoic sections of the western part of the Gornyy Altai, the Rudnyy Altai, and the Kalba. For this reason they regarded the zones of deformation separating the Irtysh and the northeastern zones of deformation as zones of profound faulting. Essentially the same conclusion as to the nature of these zones was reached by Nekhoroshev [16] in 1956. In 1953, Ivankin interpreted the Irtysh deformed zone as a special "structural-facies zone" which developed over a long period of time at the juncture between two structures, the Rudnyy Altai and Kalba.

Almost contemporaneously with these works a paper was published by Nekhoroshev [15] reviewing the structures of the entire Altai region together with those of the adjacent parts of northeastern Kazakhstan. According to the author's concept (greatly refined as compared with his earlier views) the Rudnyy Altai, as a whole, is a monocline located on the northeastern side of the Zaysan geosyncline and adjacent to the Caledonian folded structure of the Gornyy Altai. Nekhoroshev emphasized the heterogeneity of the inner structure of the Rudnyy Altai, which falls into a series of anticlinoria and synclinoria. The zones separated by him differ in stratigraphy, facies, and structure.

In 1955, based on a great amount of reconnaissance and detailed work, D.I. Gorzhevskiy, V.A. Komar, and G.F. Yakovlev published their views on the structure of the Rudnyy Altai zone and its relation to the neighboring structures [4, 5, 6]. These authors separated three "structural-facies and metallogenetic zones" within the Rudnyy Altai itself, which they interpreted as geanticlines and geosynclines: the Aley geanticline similar in outline to Nekhoroshev's Aley anticlinorium, and the Bystrushensk and Leninigorsk-Zyryanovsk geosynclines. To the southwest and northeast of these zones, lie the Irtysh and Kholzun zones extending from the Rudnyy Altai into the Kalba and the Gornyy Altai.

After an analysis of the lithology and

:ucture of the stratigraphic sections in ve various areas of the Rudnyy Altai, the thors conclude that in the Paleozoic (begin-ng with the Silurian) the Rudnyy Altai was t a graben but a "large geanticline lying ithin the Zaysan geosynclinal system." ven during that time, this geanticline was parated from the neighboring structure by e Irtysh and Kholzun zones. The same con-usions may be made from Ivankin's de-riptions (1953) of this region, although the thor himself does not draw them.

In the middle Paleozoic, beginning with e Middle Devonian, according to Gorzhev-iy, Komar, and Yakovlev [5, 6], the idnyy Altai geanticline was broken up into number of smaller structures. These ructures were named the Aley and Lenino-rsk-Zyryanovsk geanticlinal zones. Between em lies the Bystrushensk "geosynclinal ne."

The first two structures were geanticlines the second order with respect to the Rud-y Altai geanticline. They were character-ed by thinner (as much as 4,000 to 5,000) middle Paleozoic deposits, interruptions sedimentation, predominance of acid trusives over basic, low brachyanticlinal lds, considerable development of granitic cks, mainly in the Zmeinogorsk intru-ve complex, and a definite type of poly-etallic ore deposits.

The Bystrushensk zone, characterized by more complete section, greater thickness Middle Paleozoic sediments (as much as),000 m), mainly basic and intermediate trusives and their tuffs, linear short lds, relatively small distribution of gran-es of the Zmeinogorsk intrusive complex, id the absence of any significant poly-etallic ore deposits, represents an "inner eosyncline."

In the opinion of these authors [5, 6], e Kholzun and Irtysh transitional zones ossess features of geosynclinal zones such s the great thickness (more than 10,000 m) the flysch facies associated with volcanic isic rocks, and the linear folds marked by ell developed schistosity. Moreover, un-ke the structure-facies zones of the Rudnyy ltai, they contain intrusives and associated lymetallic ore deposits, mainly of the alba type.

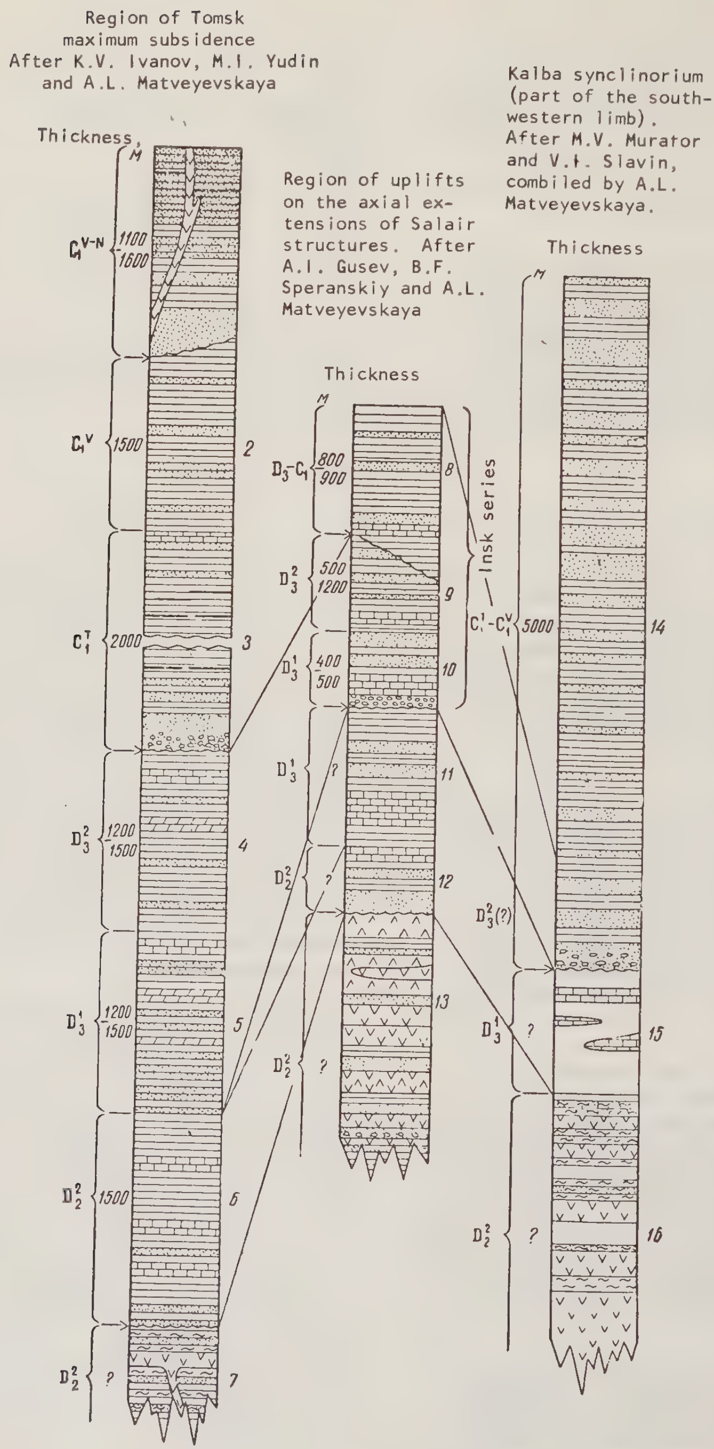
In his paper, "Paleozoic History of Rud-y Altai" [4] Gorzhevskiy also discusses e position of this block in the general ructural plan of the Sayan-Altai region. iting I. I. Belostotskiy's data on the Gornyy ltai, he denies the platform character of e latter during the Middle Paleozoic. As proof, he cites the thickness and facies composition of the Siluro-Devonian section of the Gornyy Altai, but it must be remarked that this section does not characterize the block adjacent to the Gornyy Altai but the Anuy-Chuy zone, which V. A. Kuznetsov calls a "superimposed" geosyncline [11]. Comparing it with the sections of the same age in the Kalba and the Rudnyy Altai, Gorzhevskiy concluded that in the early and middle Paleozoic, the Kalba and the Gornyy Altai had the same geologic history and were parts of a single "Zaysan geosynclinal system," while the Rudnyy Altai during the Silurian period was a region of median uplift within this geosynclinal system [4, 6]. Later, during the Devonian, in spite of the general subsidence of all three structures, their mutual relationship was preserved, in Gorzhevskiy's opinion, although the maximum subsidence occurred in the region of the Gornyy Altai.

Very similar views on the position of the Rudnyy Altai are expressed by N. I. Belostotskiy in his last paper [2]. He believes that a single "Zaysan-Altai basin extended during the Middle and Late Devonian from the Chingiz Range in the southwest to the Kuray basin and the western base of the Sumultinskiye Belki Range in the northeast."

In 1952-1953, the present author made a detailed study in the exposed areas of the Ob-Zaysan Variscan geosynclinal system. While working in its northern part adjacent to the Ob River, known as the Kolyvan'-Tomsk folded zone or arc, the author discovered that the zone has a complex structure]17, 18, 19]. Here the Ob-Zaysan Variscan geosyncline system splits into a series of positive and negative inner and marginal structures which developed over a long period of time.

Between this geosynclinal system and the bordering Kuznetsk-Alatau and Altai-Salair [18, 19] blocks, which were parts of the Variscan platform [18, 19], lie narrow but some rather wide longitudinal and lateral marginal downwarps, locally arranged en echelon. They are filled with thick (of the order of 6,000 to 11,000 m), in places much dislocated Devonian, Carboniferous, and Permian deposits. The characteristic features of these deposits are the presence of redbeds alternating with thick carbonate sediments, the presence of terrigenous deposits, and also many lacustrine-paludal, partly paralic coal-bearing deposits. In the region of the Kolyvan'-Tomsk arc, these structures are respresented by the Yel'tsov, Gorlov, Kuznetsk, and Tashmin marginal downwarps.

Over a considerable distance these marginal downwarps of the Variscan geosyn-


Region of Tomsk maximum subsidence After K.V. Ivanov, M.I. Yudin and A.L. Matveyevskaya
Region of uplifts on the axial extensions of Salair structures. After A.I. Gusev, B.F. Speranskiy and A.L. Matveyevskaya
Kalba synclinorium (part of the south-western limb). After M.V. Murator and V.I. Slavin, combiled by A.L. Matveyevskaya.
Thickness, M
Thickness
Insk series
1100
1600
1500
2000
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1500
800
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clinal system are separated from its inner part by relatively narrow geanticlines of cordillera type. These cordilleras are composed essentially of Devonian subaqueously extruded spilitekeratophyre volcanics. In the uplifted parts of the geanticlines, usually located on the continuations of the anticlinal axes of the platform, the ancient Precambrian or Lower Paleozoic (Cambrian, rarely Silurian) rocks of the same kind as the rocks of the adjacent parts of the platform outcrop directly beneath the Devonian sediments. The Mitrofanov, Bugotak, and Ordynsk structures of the Kolyvan'-Tomsk folded region are geanticlines.

To the northwest of these small structures lies the inner zone of the Ob-Zaysan Variscan geosynclinal system. It is composed of a thick series (5,000 to 10,000 m) of flyschlike black shales of Middle and Upper Devonian as well as Lower Carboniferous ages crumpled into tight linear folds. Locally, the rocks of the inner zone of the geosynclinal system and, to a less extent, of its platform are intruded by numerous granitic bodies.

Comparison of the Kolyvan'-Tomsk and the Irtysh-Zaysan (or Kalba) parts of the Variscan geosynclinal system shows that there is a great deal of resemblance between them and that some features of their development are identical.

The Kalba zone (Fig. 1) is composed of a monotonous sequence of sediments of the same age, composition, and thickness as those of the inner downwarp of the Kolyvan'-Tomsk Variscan geosynclinal system. These sediments are crumpled into tight linear folds. However, the composition and facies of the sediments in the exposed areas of the inner zone of the Variscan geosynclinal system are in most cases unlike those of the sediments of the same age in the neighboring positive and negative marginal structures.

The black shale flysch formation of the Kalba zone, as well as of the Kolyvan'-Tomsk zone, is intruded by numerous bodies of Variscan granites. Many of the composition characteristics and age and genetic relationships of these granites known in the Irtysh-Zaysan part of the Variscan geosynclinal system, are observed also in its Kolyvan'-Tomsk zone.

The submerged part of this zone, which connects the Kolyvan'-Tomsk and Kalba sections, is clearly traceable on magnetic anomaly graphs. It has the shape of an arc, concave to the east-southeast. According to the airborne magnetic survey by the Siberian Geophysical Organization, the submerged section of the inner downwarp of the Ob-Zaysan geosynclinal system is characterized mainly by a negative or

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FIGURE 1. Stratigraphic sections of the main downwarp of the Ob-Yenisey Variscan geosynclinal system

1 -- Gray sandstones with subordinate dark-gray siltstones and shales containing layers of coal (upper Visean shale-sandstone sequence);
2 -- Dark-gray shales with siltstone beds (Lagerny Sad beds);
3 -- Dark-gray argillaceous, in places marly and carbonaceous shales with some beds of siltstones and sandstones; sandstones and silstones.
4 -- Dark-gray and greenish shales with a few beds of siltstones, marly shales, and limestones (Upper Devonian sequence);
5 -- Calcareous sandstones and marly shales; gray and green chloritized, commonly coarse-grained micaceous sandstones with layers of black shales;
6 -- Gray, mainly argillaceous and marly shales with subordinate beds of sandstones and arenaceous-argillaceous limestones; some beds of lilac and ferruginous shales and sandstones;
7 -- Volcanics, mainly basic and intermediate: breccias, tuff breccias, and sandy tuffs; Inskaya series;
8 -- Fossiliferous calcareous shales with thin beds of fine-grained sandstones; the shales pass into arenaceous and marly sediments (Schiefer beds);
9 -- Dark-gray calcareous shales and sandstones; lenticular limestone layers at the base (spore-bearing beds);
10 -- Gray micaceous sandstones with flora and some layers of shales; limestones, conglomerates, sandstones, and shales (Iniodendron beds);
11 -- Green-gray, green, yellow, and brown shales and sandstones passing downwards in the section into coral-bearing limestones, sandstones, and red-brown and green shales;
12 -- Dark-gray limestones, bluish-gray and green shales, and sandstones;
13 -- Basic and intermediate volcanics with rare beds of sedimentary rocks, and acid sills; acid extrusives predominate in the upper part (Bugotak formation);
14 -- Gray and green sandstones, partly sandy tuffs, alternating with argillites and shales; in the lower part of the section, dark-gray and gray shales with sandstone beds predominate; the basal beds are composed of conglomerates and sandstones (Kalba formation);
15 -- Green, red, and violet argillites, siliceous argillites and cherts with lenses and beds of limestones in the upper part of the section;
16 -- Green tuffs, sandy tuffs, and argillites interbedded with andesite and basic flows.

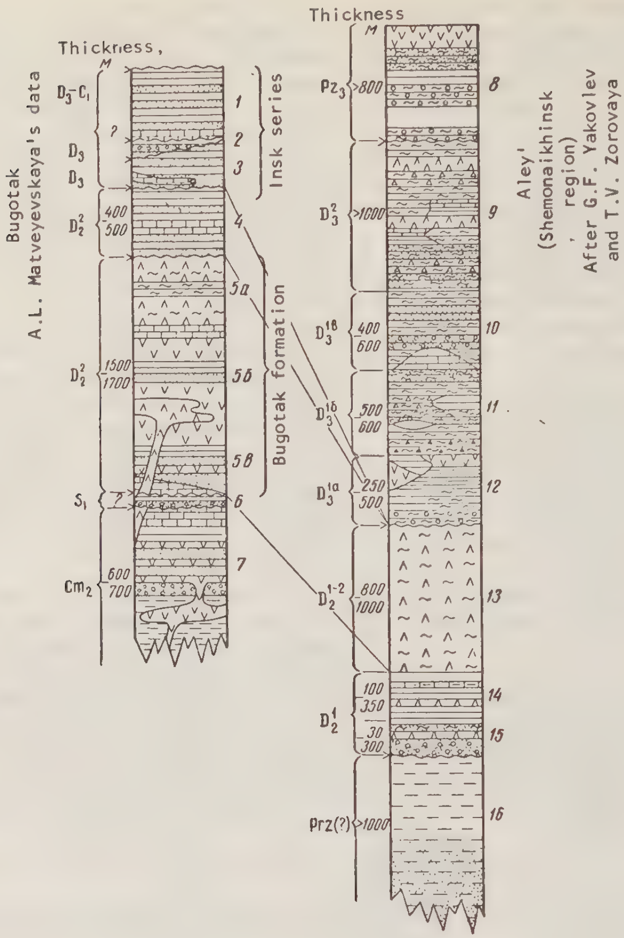


FIGURE 2. Stratigraphic sections of the geanticlinal areas of the Ob-Yenisey Variscan geosynclinal system

1, 2, 3 -- Insk series:

1 -- Calcareous shales with thin beds of fine-grained sandstones (Schiefer layers);

2 -- Gray calcareous shales, sandstones (spore-bearing beds);

3 -- Gray micaceous sandstones and shales (Iniodendron beds);

4 -- Red-brown, green, yellow, and brown shales and sandstones, limestones, gray sandstones, and shales;

5 -- Bugotak formation: 5a -- mainly acid extrusives, 5b -- basic and intermediate extrusives with a few beds of tuffaceous and sedimentary rocks and acid sills; thin-bedded unfossiliferous limestones;

6 -- Shales, sandstones, conglomeratic sandstones;

7 -- Marbles, schists, sandstones, and conglomerates; strongly metamorphosed and enclosing numerous sills, schists (Ikovsk formation);

8 -- Sandstones with layers of carbonaceous shales carrying a flora, siltstones, sandy tuffs, and conglomeratic tuffs; andesites, acid, and intermediate tuffs;

9 -- Acid extrusives with their tuffs and breccias, tuffaceous sediments, lenses of siltstone, shales, and limestones;

10 -- Acid tuffs, conglomerates, polymict sandstones, lenses of limestone at the base.

11 -- Coarse-grained acid tuffs, sandy tuffs, tuff breccias, siliceous shales;

12 -- Argillaceous-siliceous and siliceous shales; acid, intermediate, and basic tuffs; tuff conglomerates, andesites;

13 -- Acid extrusives and their tuffs and breccias;

14 -- Argillites, calcareous-argillaceous, calcareous-siliceous shales, sandstones, limestones, acid extrusives, and tuffs;

15 -- Conglomerates, sandstones, argillites, sandy tuffs, tuffaceous sediments, acid extrusives, and tuffs;

16 -- Schists and metamorphosed sandstones.

weakly positive undisturbed magnetic field similar to that of the exposed areas of the inner zone of the Variscan geosynclinal system. Some of the areas of this section, however, have a few small and rather intensive positive anomalies corresponding, apparently, to the intrusive bodies.

In comparing the Aley zone of the Rudnyy Altai, which has a definitely geanticlinal character, with the structures of the Kolyvan'-Tomsk arc, we note that in its composition and development the Aley zone is very similar to the Ordynsk, Bugotak, and Mitrofanov geanticlines (Fig. 2), for in both places the base of the Middle Paleozoic is underlain by the Lower Paleozoic and possibly by the Precambrian sediments of the same type as in the adjacent parts of the Variscan platform. The recently published data on the similarity between the Rudnyy Altai schists and the Terekty metamorphic complex of the Charysh-Terekty massif confirm the facts observed in the region of the Kolyvan'-Tomsk arc.

Ordovician and Silurian deposits, with the exception of some small exposures of the latter in the Ordynsk geanticline which are analogous to the platform deposits of Salair, are unknown. Devonian sediments in all investigated structures begin with volcanic-sedimentary Middle Devonian deposits. If the presence of Lower Devonian beds in the Rudnyy Altai is finally proved, then the formation of the southern part of the Variscan geosynclinal system must have occurred somewhat earlier. The Middle Devonian deposits of the Rudnyy Altai and of the geanticlines of the Kolyvan'-Tomsk arc are identical in composition, position in the section, degree of diagenesis, and type of folded structures. The Upper Devonian deposits in the geanticlines of the "second order" in the Rudnyy Altai are unknown in the Kolyvan'-Tomsk zone, but this is the result, perhaps, of insufficient knowledge of the stratigraphy of the volcanic formation in the latter region. The younger formations, including the upper Variscan granites, are also unknown in the Ordynsk, Bugotak, and Mitrofanov geanticlines. However, in the Rudnyy Altai also, Carboniferous and younger sediments accumulated only in the negative or the "inner geosyncline" zones as they are called by the students of the Rudnyy Altai.

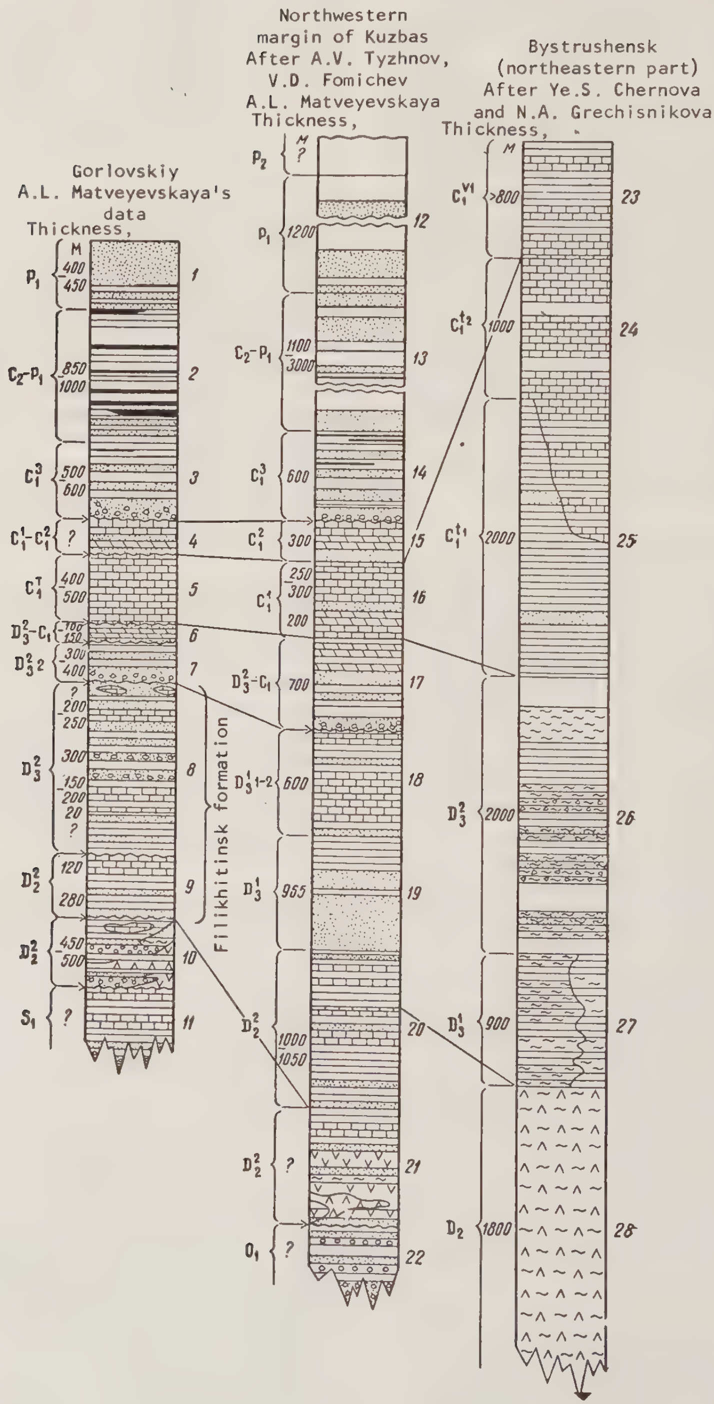
Thus, the close resemblance of the geanticlinal structures of the Kolyvan'-Tomsk part of the Variscan geosynclinal system to the geanticlinal structures of the Rudnyy Altai is quite obvious. It is possible that the latter developed over a longer period of time and this may be the reason for their greater size.

The present author cannot agree with Gorzhevskiy [4] and Belostotskiy [2], who deny the platform character of development of the Gornyy Altai during the Middle Paleozoic. In presenting Devonian sections, Gorzhevskiy does not consider the heterogeneous structure of the Gornyy Altai. An analysis of stratigraphic sections of individual zones and their correlation with those of the neighboring regions, according to V.A. Kuznetsov, Belostotskiy himself, and other investigators of the Altai Mountains, shows that beginning with the Silurian the blocks adjacent to the Rudnyy Altai (Charysh-Terekty and Talitsk, Figs. 3 and 4) and the older Altai-Kuznetsk folded and block faulted zone [2, 11] of eastern Altai were positive structures with erosion predominating over sedimentation. These blocks represent two superimposed structures. The lower structure consists of tight linear folds in Precambrian and Lower Paleozoic rocks. The Silurian and Middle Devonian sequences of the upper structure, found only in some localities, are relatively thin, (of the order of 3,000 to 4,000 m) form folds of the platform type, and are much less faulted.

The accumulation of thick, mainly Middle and Upper Paleozoic sediments described by Belostotskiy from the Gornyy Altai and cited by Gorzhevskiy to prove his view, occurs only in the narrow zone of the Anuy-Chuy "superimposed geosyncline" [12]. Thus, the geanticlinal or essentially platform character of development of the eastern and western Salair-Caledonian blocks of the Gornyy Altai during the Middle and Upper Paleozoic cannot be denied after comparison of the sections.

Even so far as the Anuy-Chuy downwarp is concerned, the section of which Gorzhevskiy used to prove his constructions, the assumption that its geological history was identical with that of the Irtysh-Zayson zone cannot be considered as proved, nor especially the statement that it underwent greater subsidence than the latter.

The Devonian sediments in these regions are very similar. But the comparisons of these regional sediment thicknesses, used by Gorzhevskiy to prove his view, cannot be accepted because the thickness given for the Devonian strata in the Gornyy Altai, is for a downwarped region (superimposed geosyncline), whereas the thicknesses of the same strata in the Rudnyy Altai and especially in the Kalba were measured in the geanticlinal areas with interrupted and much reduced sections. It is difficult to say what the thicknesses of the Devonian deposits (mainly Lower and Middle Devonian) are in the downwarps of the Irtysh-Zaysan Variscan geosyncline. It is likely that they increase


Northwestern
margin of Kuzbas
After A.V. Tyzhnov,
V.D. Fomichev
A.L. Matveyevskaya
Thickness,
Bystrushensk
(northeastern part)
After Ye.S. Chernova
and N.A. Grechisnikova
Thickness,
Gorlovskiy
A.L. Matveyevskaya's
data
Thickness,
Filikhitinsk formation

considerably in these areas, especially because, according to a communication from Gorzhevskiy himself, the subsidence of the pre-Paleozoic basement which, according to the geophysical data, is nearest to the modern surface within the boundaries of the Rudnyy Altai, occurred gradually in the direction of the Gornyy Altai (evidently the Talitsk block) and more intensively towards the Kalba.

Thus, the development not only of the Gornyy Altai as a whole, but also of its Anuy-Chuy zone and of the inner zone of the Irtysh-Zaysan geosynclinal system during the Early and Middle Devonian, was similar but by no means identical. Beginning in the Late Devonian, the downwarping of these regions was very different and they cannot be considered as a single structure ("Zaysan" according to Gorzhevskiy). This is confirmed by the data of airborne magnetic surveys in the regions adjacent to the front of the Altai Mountains. As has already been stated, these data quite clearly reveal the unity of structures in the Irtysh-Zaysan and Kolyvan'-Tomsk geosynclinal systems, but do not indicate that these structures continued into the Gornyy Altai.

Returning to the structures of the Rudnyy Altai itself, we must note that, having insufficient data at present on all the small structures of the Rudnyy Altai zone found here by Rudnyy Altai geologists [6], we cannot make a sufficiently exact geotectonic subdivision of the region. However, it is possible at present to construct a working outline of the regional structure on the basis of general concepts of the structure of the Ob-Zaysan Variscan geosynclinal system and of the bordering platform (Fig. 5) derived from our investigations. There is no doubt that the Aley anticlinorium of the Rudnyy Altai is a geanticline of the Variscan geosynclinal system similar to those of the Kolyvan'-Tomsk arc.

The Leninogorsk-Zyryanovsk zone is either a small geanticline en echelon with the Aley geanticline, or it may belong to the western Caledonian structure of the

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FIGURE 3. Stratigraphic sections of the marginal downwarps of the Ob-Yenisey Variscan geosynclinal system

1 -- Green sandstones with beds of shale and thin layers of coal (Kuznetsk formation);
2 -- Greenish-gray sandstones and shales with coal beds (Balakhonsk formation);
3 -- Quartzose sandstones, gray calcareous shales, basal conglomerates - puddingstones (Yelbashinsk formation);
4 -- Yellow-gray sandstones and marly shales (Marly formation);
5 -- Gray reef limestones, fossiliferous (Lower Limestone formation);
6 -- Marly shales, sandstones, limestones;
7 -- Red-brown shales, conglomerates, and sandstones;
8 -- Yellow-gray, green-gray and brown shales and sandstones with lenses and layers of limestones, red-brown sandstones, and shales;
9 -- Dark-gray limestones, bluish-gray and green shales with sandstone beds;
10 -- Conglomerates, sandstones, and shales with a few acid flows (Bugotak formation);
11 -- Sandstone - shale cyclothem with white crinoidal limestones (Legostayev formation);
12 -- Yellow-gray, gray sandstones, dark-green siltstones and argillites with beds of marlstones;
13 -- Gray sandstones, arenaceous-argillaceous, argillaceous, and carbonaceous shales with numerous thick beds of coal (Balakhonsk formation);
14 -- Cyclothem of conglomerates, strong quartzose sandstones and siltstones with thin beds of coal (Ostrog formation);
15 -- Gray, dark-gray limestones, light-gray, green and pink sandstones and marlstones;
16 -- Dark-gray, gray, in places arenaceous silicified limestones with siliceous concretions, dolomites (Lower Limestone and Abyshev formations);
17 -- Red-brown and green sandstones, marly shales, and marlstones (Upper Redbeds);
18 -- Green shales with beds of arenaceous limestones; black, some pink and light-gray limestones; green-gray shales with sandstone beds (Glubokinsk formation);
19 -- Light-gray, fine-grained sandstones, bluish-gray shales, arenaceous limestones;
20 -- Greenish-gray shales with limestone beds, gray arenaceous limestones (Zarubinsk formation;
21 -- Bluish-gray, less commonly greenish phyllitized shales with beds of sandstones and limestones (Yashkinsk and Vlaskov limestones); argillaceous and arenaceous shales; sandstones, commonly tuffaceous; conglomerates, tuff breccias, diabases, quartz porphyries and their tuffs;
22 -- Green and violet sandstones and shales;
23 -- Interbedded crinoidal limestones, calcareous shales, and shales (Ul'binsk formation);
24 -- Crinoidal limestones, calcareous-argillaceous shales, argillites (Bukhtarminsk formation);
25 -- Argillaceous shales with some siltstones; in the northeast and in the upper part of the section, calcareous argillites and argillaceous limestones with beds of siltstones (Tarkhansk formation);
26 -- Argillites, siltstones, tuffaceous sandstones, tuffs and tuff breccias of siliceous composition;
27 -- Acid, intermediate, and basic tuffs with beds of argillites and tuffaceous sandstones; in the southeast, argillaceous shales with layers of acid tuffs;
28 -- Acid extrusives, their tuffs and tuff breccias.

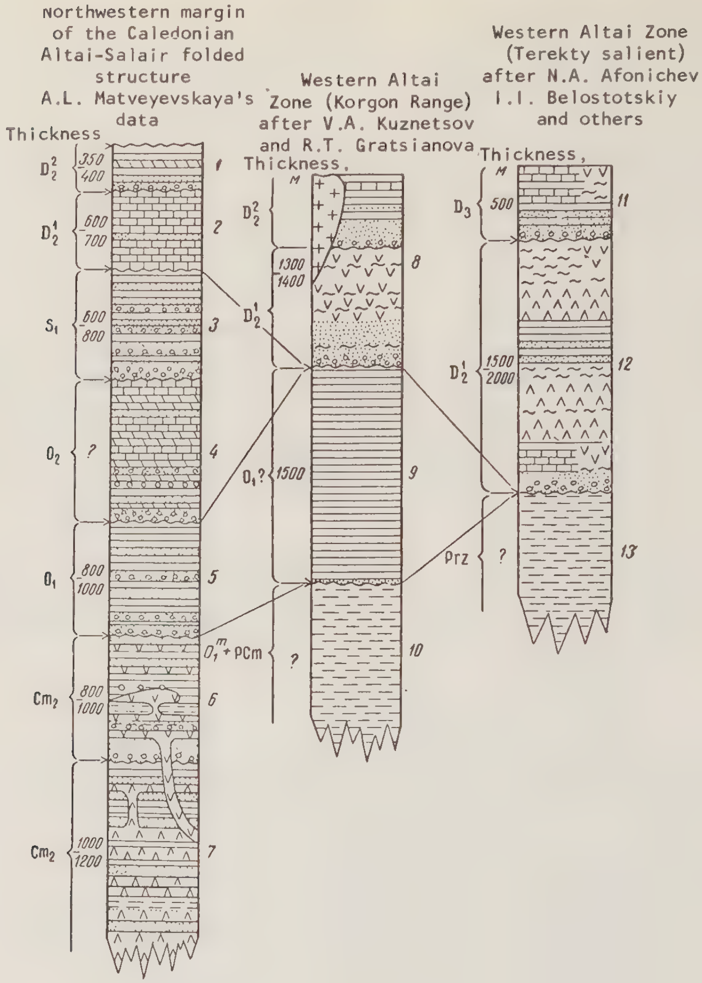

FIGURE 4. Stratigraphic sections of the Altai-Salair Variscan (epi-Caledonian platform)

1 -- Sandstones and shales with thin beds of limestones (Khmelev formation);
2 -- Dark-gray reef limestones with beds of sandstones and shales (Sokolinogorsk formation);
3 -- Cyclothemic sequence of sandstones and shales with thin beds of white crinoidal limestones; conglomerates and sandstones at the base (Legostayev formation);
4 -- Conglomerates, sandstones, thin-bedded marlstones, and limestones;
5 -- Green and violet conglomerates, sandstones, and shales (Green-Violet formation);
6 -- Shales, sandstones, and conglomerates, strongly metamorphosed and containing numerous sills and flows of basic and intermediate composition (Ikovsk formation;
7 -- Sedimentary - tuffaceous rocks (schists) with subordinate acid flow; in the middle part of the section, quartz albitophyres predominate, schists are subordinate;
8 -- Upper part: gray sandstones, shales, and limestones (upper subformation of R.T. Gratsianova); lower part: abundant acid flows and their tuffs underlain by basal conglomerates and sandstones (Korgon formation);
9 -- Shales (Green-Violet formation
10 -- Schists;
11 -- Intermediate and basic extrusives passing laterally into sandstones, shales, and limestones;
12 -- Mainly acid extrusives, their tuffs, green and lilac sandstones, and shales;
13 -- Schists (Terekty complex).

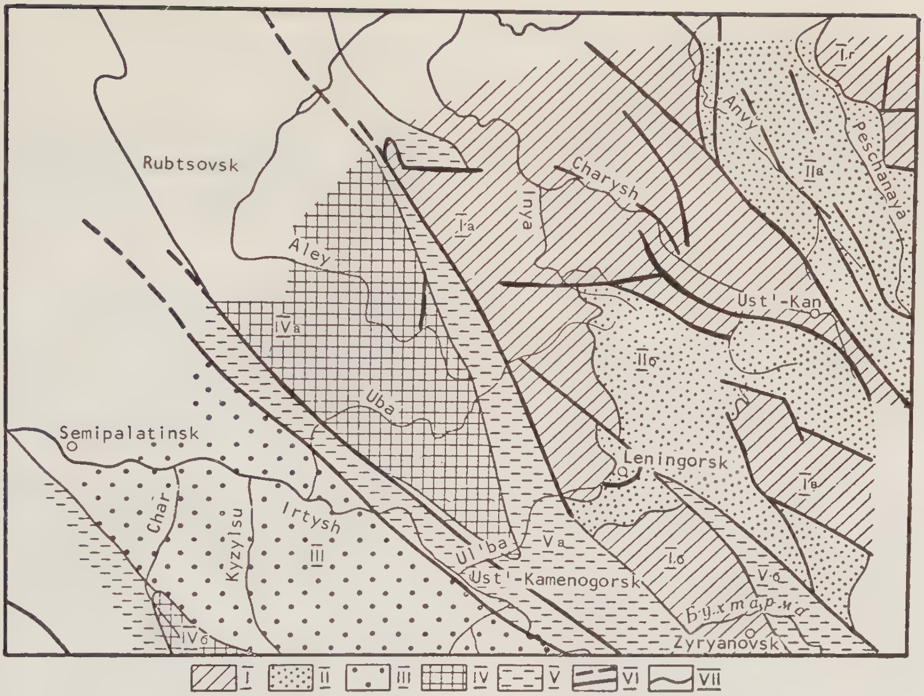

FIGURE 5. Structure of the Rudnyy Altai and adjacent areas

Variscan epi-Caledonian platform:
I -- anticlinal zones (a- Talitsk, b- Lenininogorsk-Zyryanovsk, Terekty, c- East-Altai);
II -- downwarps (a- Annuy-Chuy superimposed geosyncline, b- Tigiretsk-Korgon)

Variscan geosynclinal system:
III -- Irtysh-Zaysan inner downwarp,
IV -- geanticlines (a- Aley, b- Char),
V -- marginal downwarps (a- Bystrushensk, b- Kholzun),
VI -- disjunctive dislocations, observed and assumed,
VII -- outlines of structures.

Gornyy Altai. For the sake of convenience, we shall call it the West Altai geanticline and include in it the Talitsk, Leninogorsk-Zyryanovsk and Charysh-Terekty blocks, together with the Tigiretsk-Korgon syncline located on them, which is transverse to the Caldeonian structures and is filled mainly with Devonian sediments. This syncline lies in the region of juncture of the blocks, and in the southwest joins the downwarped area of the Rudnyy Altai (Kholzun zone).

The Bystrushensk zone, which includes the Zmeinogorsk, Bystrushensk, and Buk tarminsk synclinoria with thick (as much as 14 km) middle and upper Paleozoic, mainly sedimentary deposits containing thick limestone sequences, together with the southern (Narymsk) area of the Rudnyy Altai and evidently the Kholzun zone of the authors, is a complex synclinorium which in the southeastern part of the Rudnyy Altai separates the Caldonides of the Western Altai from the inner zone of the Ob-Zaysan Variscan geosynclinal system. In the northwestern Rudnyy Altai it lies between the first and the Aley geanticlines. It is separated from the inner zone of the Variscan geosynclinal system by the Irtysh deformed zone which, farther on, separates this inner zone from the Aley geanticline. The northwestern continuation of the synclinorium lies, evidently, near the village of Kur'i, where upper Tournaisian sediments lie transgressively on the Devonian and older strata.

Judging by the completeness of its section, the age of its sediments, their thickness, the character of the facies and their position, this synclinorium should be correlated with the transverse marginal downwarps of the Kolyvan'-Tomsk folded zone (Fig. 3).

In spite of all these resemblances this marginal downwarp possesses some distinguishing features. The most important among them is the presence of the upper Paleozoic acid extrusives and of a large number of Upper Variscan intrusives (mainly of the Kalba type).

The Irtysh deformed zone is a characteristic feature of the marginal structures of the Variscan geosyncline related, apparently, to a very large and deep, faulted zone.

As for the Irtysh-Zaysan part of the geosynclinal system as a whole, the above data show that it (and not only the Rudnyy Altai part of it, as is believed by Belostotskiy [2]), as well as the analogous structure of the Kolyvan'-Tomsk arc, had its beginning in the Early Devonian on the Caledonian folded basement, whose rocks outcrop in the neighboring regions of the Variscan platform of Kazakhstan, in the Altai-Salair region, and in the geanticlinal structures of the geosynclinal system itself. Only from this point of view is it possible to agree with Gorzhevskiy's opinion that during the early and, in part, middle Paleozoic time, the Irtysh-Zaysan and the Gornyy Altai regions had a "common geological history."

Thus the materials on the structure and history of the development of the Rudnyy Altai itself and of the neighboring structures of the Ob-Zaysan Variscan geosynclinal system show that the Rudnyy Altai is a region of small fault-block structures at the zone of juncture between the Variscan geosynclinal system and the epi-Caledonian Gornyy Altai platform which borders it in the east.

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ON THE RELATION BETWEEN DEPOSITION OF SULFUR AND FRACTURE TECTONICS

by

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Of the various types of sulfur deposits, the most important economically are sedimentary deposits occurring in beds and lenses localized in definite stratigraphic zones. The characteristic features of such deposits are: the association of sulfur with bitumens or other hydrocarbons, calcite, gypsum, and anhydrite and its accumulation, mainly in limestones (or dolomites), less commonly in marlstones and other rocks, near their contact with gypsum and anhydrite, the latter feature being the reason for the stratigraphic localization of sulfur deposits. Such deposits occur on platforms (the Volga group), in folded regions (Fergana, Gaurdak, Sicily) or on the boundary between a platform and a region of deep subsidence (Western Ukraine). Although the enclosing rocks may be of any age, the morphology of the ore bodies and the composition of the associated minerals alike, and the geologists who have studied their origin refer them to a single genetic group. Therefore, the discovery of characteristic features of any one of these deposits is significant for the study of other deposits of this type.

The numerous hypotheses for the genesis of sedimentary sulfur deposits fall into two groups: according to one, sulfur is syngenetic; to the other, epigenetic. The theoretical and practical importance of the problem of the origin of sulfur is obvious.

Even without reviewing the arguments concerning the mechanism of formation of sulfur deposits, it is evident that tectonics must play an important role in the epigenetic hypotheses. A.S. Uklonskiy [8] pointed out that the porosity of rocks and fractures of all types aids in sulfur accumulation, which in his opinion is the product of hydrogen sulfide oxidation. Later, A.S. Sokolov [7] noted the localization of sulfur deposits in definite geologic structures, especially in the positive ones such as the salt domes, anticlines, and brachyanticlines of the folded regions and in the low domical uplifts and certain monoclines of the platforms. He drew attention to the greater accumulations of sulfur in the crests of folds and on those limbs of folds and sometimes of monoclines which are turned towards the broadest and deepest synclinal downwarps.

N.P. Petrov [5] considers that one of the main factors in the formation of the Gaurdak sulfur deposit is the intersection of a fold with a large tectonic dislocation, the Usun-Kuduk graben. Observations at the Gaurdak locality showed a definite relation between sulfur deposition and the tectonic features of the dome.

The Gaurdak sulfur deposit lies in the southwestern end of the dome, which is a part of one of the western virgations of the Gissar Range. The deposit is in the limestone-anhydrite sequence (lower) of the Gaurdak formation (Kimmeridgian-Tithonian). This sequence is 150 to 50 m or less in thickness (75 to 80 m on the average) at the deposit and is composed of anhydrites partially changed to gypsum, which contain lenses and beds of limestone ranging in thickness from one to several tens of meters and restricted to three nonpersistent beds. It is noteworthy that, upwards in the section, as the limestones become thinner they also become more persistent. The uppermost limestone bed, which separates the anhydrite-limestone sequence from the overlying anhydrite beds, is called the "key bed" because of its wide distribution. The ore bodies are localized in three levels corresponding to three limestone layers and are most extensive in the lowermost of "F" layer, less extensive in the middle of "D" layer and least extensive in the key bed. The productive beds are underlain by the Kugitang formation (Lusitanian), which is composed of thick-bedded dark-gray and black, dense, very strong bituminous limestones. The thickness of this formation is 735 m. There is an unconformity between the Garudak and the Kugitang formations.

The thickness of the higher anhydrite

and salt-bearing beds of the Gaurdak formation near the dome is about 800 m. At the sulfur deposit the salt-bearing beds are absent and the anhydrites do not exceed 150 to 200 m in thickness. The reasons for this will be discussed below.

The Gaurdak formation in the dome is overlain with a pronounced angular unconformity by Lower and locally Upper Cretaceous sandstones. The thickness of the Lower Cretaceous beds within the sulfur deposit is 45 to 500 m, and of the Upper Cretaceous beds, 1500 m. The Upper Cretaceous sediments are overlain by 144 m of the Bukhara sediments (Paleocene) composed of gypsiferous sandstones and gypsum in the lower part (50 m) and of foraminiferal limestones (94 m) higher in the section. The younger sediments of the region are Quaternary fanglomerates 200 m in thickness. The Gaurdak dome is an uplifted part of the Gaurdak-Tyubegatan anticline, which parallels the Kugitang anticline (Fig. 1). The downwarp between these anticlines contains Upper Cretaceous and, locally, Paleocene (Bukhara stage) sediments. The domical uplift, amounting to about 2,500 m, includes rocks of the Paleozoic crystalline basement and Jurassic and Lower Cretaceous strata, considerably complicated by salt tectonics, which produced a number of small domes that partially surround the Gaurdak dome in the south. In the southwest and southeast, the strata composing the dome are cut by two large faults, the Uzun-Kuduk graben and a hinge fault (?) striking to the northeast. The displacement of these faults amounts to 1,000 to 2,000 m.

The sulfur deposit lies in the periclinal end of the dome. The Kugitang formation, which forms the core of the dome, dips to the southwest, west, and northwest within the boundaries of the deposit. The dip changes from 16° to 30°, increasing noticeably from the center to the periphery. No large faults have so far been found within the core, but horizontal and normal faults with displacements from a few to 20 m have been noticed in different parts of the deposit; for example, northwest of the shaft of the active mine there is a fault with a displacement of about 25 m striking 290° to 300° and dipping 200° to 210°[1] at 75° to 85°.

The unconformity between the Gaurdak and the Kugitang formations is revealed only by examining their contact over a large area. In the northeastern part of the dome the Gaurdak formation lies on the

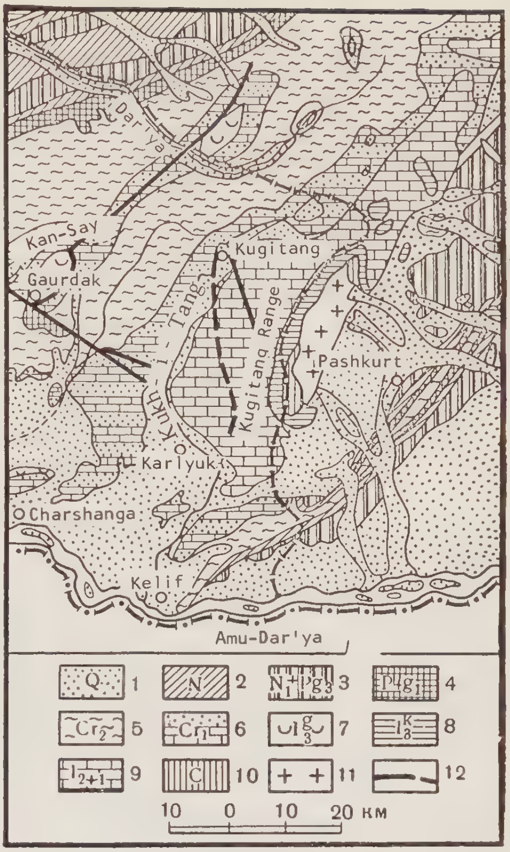

FIGURE 1. Geologic map of the Gaurdak-Kugitang region.

1 -- Quaternary sediments. 2 -- Neogene, undifferentiated. 3 -- Lower Neogene, undifferentiated. 4 -- Paleocene, Bukhara formation. 5 -- Upper Cretaceous. 6 -- Lower Cretaceous. 7 -- Gaurdak formation, Kimmeridgian-Tithonian. 8 -- Kugitang formation (Callovian-Oxfordian). 9 -- Coal-bearing sediments, undifferentiated, lower and middle sections. 10 -- Paleozoic deposits, undifferentiated. 11 -- Variscan granites. 12 -- Faults.

lower beds of the Kugitang formation. The thickness of the section is about 100 m. Considering the normal marine character of the sediments of the Kugitang formation, and their persistence over a long distance, the assumption that this unconformity is due to facies change is improbable.

Within the area of the deposit the general attitude of the limestone-anhydrite or sulfur-bearing sequence analogous to the attitude of the Kugitang formation (Fig. 2). It strikes and dips in the same direction with the angle of dip ranging from 15° to 22°. The smaller angle of dip as compared with that of the Kugitang formation is explained by the increase in the thickness of the sulfur-bearing beds from the center

[1]Strikes and dips are given in azimuths measured from the north.

to the periphery of the dome, as can be clearly seen in the section (Fig. 2). In the region of the Karacha canyon, the key bed lies almost directly on the Kugitang formation; i.e., the thickness of the sulfur-bearing beds diminished to 10 to 20 m, and possibly to less than that. The decrease in the thickness of the Gaurdak formation towards the center of the dome affects not only the sulfur-bearing beds but also the higher beds. This is confirmed by the discovery of the salt-bearing beds, the thickness of the middle, anhydrite, part of the Gaurdak formation must be one half or one third as thick as in the normal section described by N.P. Petrov. As for the thickness of the salt, very possibly it has been diminished by being squeezed out.

Joints are developed everywhere in the rocks of the Kugitang and Gaurdak formations, but with varied intensity. In order to determine their relation to structure, they were studied in different parts of the dome, located in the second, third, and fourth exploratory areas. In each are the joints in the sulfur-bearing beds and in the underlying Kugitang formation were studied separately. Bedding joints were disregarded. The joints in the Kugitang formation are developed as follows:

a) The Second Area. The Kugitang limestones, dipping 205° SE at an angle of 20°, outcrop near the northeastern boundary; in Fig. 3 their attitude is shown by the arcs of great circles. Because of their persis-

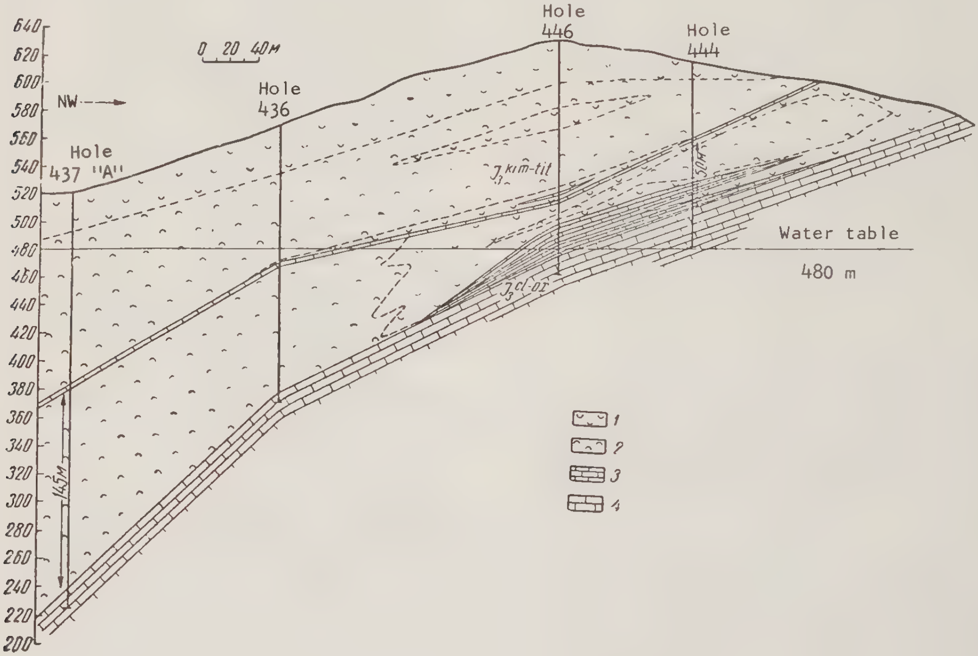

FIGURE 2. Geological section of the sulfur deposit.

Data of the Gaurdak G.R.E., geochemical prospecting. 1 -- Gypsums. 2 -- Anhydrites. 3 -- Gaurdak limestones. 4 -- Kugitang limestones.

The character of folding is best revealed by the attitude of the key bed and the limestones "D" and "F" both in boreholes and outcrops. The strata are crumpled into longitudinal and transverse folds with widths of 10 to 300 m and amplitudes of 10 to 20 m. The steepness of the folds decreases from the periphery to the center but their width increases.

Normal, horizontal, and other faults are very difficult to locate in the anhydrites, and then mainly by indirect evidence.

tence, weathering, the presence of canyons, and clear-cut bedding, the joint systems are clearly seen both in sections and on the bedding surfaces. In the latter case the joints filled with white coarse-grained calcite stand out particularly well against the black color of the rocks. The results of plotting the joints on circular diagrams are given in Fig. 3. The diagram for the second area shows two distinct maxima with joint density of 8 to 10 percent. The first maximum (Ia) is due to steeply dipping northwest trending joints with average strike of 311°. These joints are either closed or

actions of a millimeter wide, usually with raight smooth walls. The larger joints, th widths of a few millimeters, are cov-ed by crusts of coarse-grained white cal-te. The joints are usually quite persistent. visible on the bedding surfaces and have smooth, straight walls. In the fourth area, along the northern wall of the Karacha canyon, this joint system coincides with a celestite vein striking 305° W and dipping

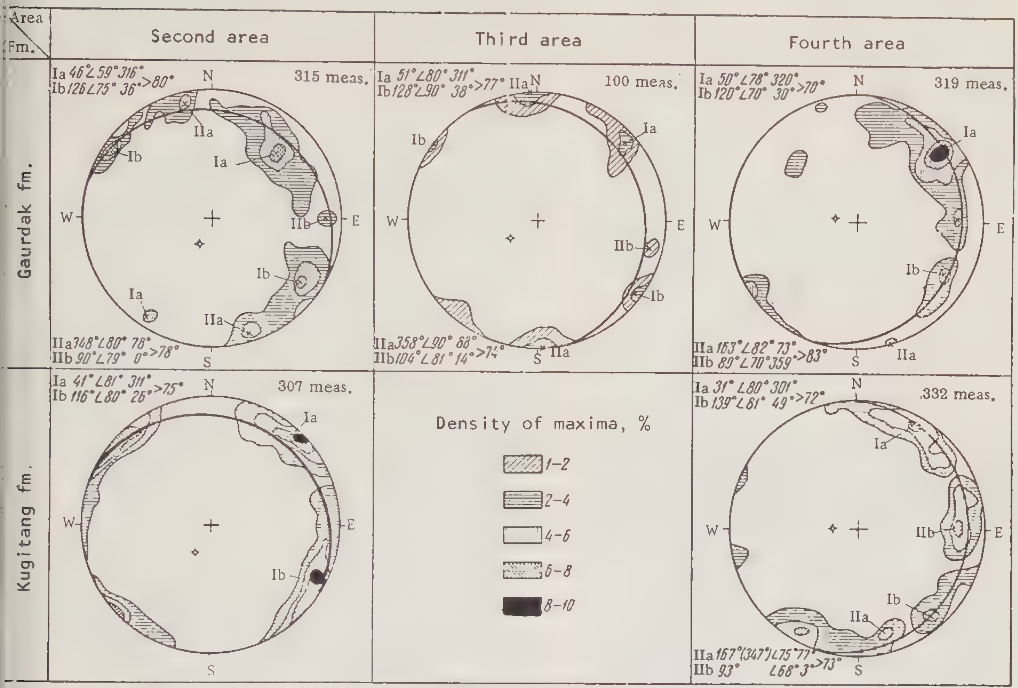

FIGURE 3. Joint diagrams.

Great circle arcs give strike of the rocks; dip angles are shown in the corners of the diagrams. Maxima of bedding joints are not shown. Because of insufficient number of measurements, the 1 to 2 percent density of maxima is shown in the third area only.

No less characteristic of this system are ones of cleavage in the limestones. Such ones, as much as 1.5 to 2.0 m in width, onsist of fine parallel cracks striking 300° 320° NW. In places the joints of this ystem, especially those dipping southwest, ave vertical slickensides on the calcite ncrusted walls.

The second maximum (Ib) is due to al-nost vertical northeast trending joints strik-6° E. Usually they are thin, almost hair-ke cracks, commonly filled with transpar-nt calcite. The wider joints with this trend requently have irregular walls.

b) The joint diagram of the fourth area also given in Fig. 3. Here also there re two maxima, one of which (Ia) char-cterizes the joints dipping 31° NE at an ngle of 80° (strike, 30°) and coincides ather well with the maximum designated the same letter in the preceding dia-ram. The morphology of the joints is milar to that in the second area; they are

75° NE. On both sides of the vein the limestone is cut by a system of thin parallel joints. Two other maxima (IIa and IIb) are due to two joint systems, one (IIa) with a sublatitudinal and the other (IIb) with a submeridional trend. The joints of the first system are rather far apart but rather wide and persistent; the joints of the second system are very closely spaced, thin, and regular.

Observations on the density of jointing in the Kugitang limestones reveal a definite relationship between the thickness of a bed and the number of joints (Fig. 4). There are from 4 to 17 joints per linear meter, or 12 on the average.

Joints in the Gaurdak formation. The gypsums weather much more readily than the Kugitang limestones and study of jointing in them is much more difficult. However, well-developed joints can often be seen in the walls of gullies and canyons and in drill hole platforms. Measurements

of the joints were made in the gypsums of the upper part of the limestone-anhydrite sequence and in the lower parts of the anhydrite beds.

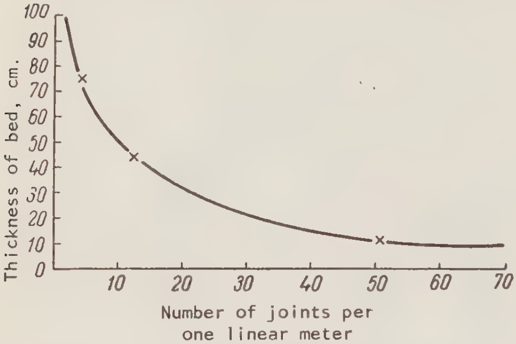

FIGURE 4. Relation between the thickness of a bed and the number of joints per meter of length. Field counts are marked by crosses.

a) The Second Area. The results of measurements are shown in Fig. 3. There are two sharp maxima -- Ia and Ib, with the joints striking 316° W and 36° E and dipping 59° and 75°, respectively. The first maximum is due to steeply dipping thin linear joints. Among the joints of the second maximum (Ib), which in general are similar to the preceding ones, joints ranging in width from 1 to 3 cm are common; they are persistent and are filled with white coarse-grained unconsolidated gypsum. Sometimes the joint walls are irregular. The other two maxima are much more weakly developed and characterize two systems of steeply dipping joints with submeridional and sublatitudinal trends. These joints are not as common as the others, but many of these are large and persistent.

b) One hundred measurements were made in the third area. They were made in an old cut near the shaft of mine No. 2. Two maxima, corresponding to joints of the Ia system (strike 311°), Ib (strike 38°) and IIa (strike 88°) appear in Fig. 3. The fourth maximum (IIb), corresponding to the nearly north-south joints, is weak.

c) The measurements of joints in the fourth area also gave four maxima corresponding to the four joint systems. As in the preceding cases, the best developed joints are those striking northwest (40°) (Ia) with an average dip of 78°. Somewhat weaker is the IIb maximum, corresponding to the submeridional joints with 70° dips. The sublatitudinal joints are poorly developed.

A comparison of jointing in the Kugitang and Gaurdak formations in the second and fourth areas shows that in the second area, two joint systems, one striking northwest and the other northeast, pass from one formation into the other almost without change in direction. The observed deviation amounts to 5 to 10°. The dips of the joints decrease slightly in passing from the Kugitang to the Gaurdak formation. The angle between the two systems is constant and equals 80°.

The submeridional and sublatitudinal joint systems, which give distinct maxima in the Gaurdak formation, do not show on the Kugitang diagram, probably because of an insufficient number of measurements.

In the fourth area, the northwest and the northeast joint systems are equally well developed in both formations, but the change in their orientation here is greater and reaches 19° to 20°, whereas the dips either remain almost the same (Ia) or, as in the second area, increase in the Kugitang formation. The angles between the joint systems change slightly, from 70° in the Gaurdak formation to 81° in the Kugitang formation.

The joint systems of submeridional and sublatitudinal trends are equally well developed in both formations, and the deviation in strike and dip is either absent or very small (4°).

This comparison of the character of jointing in the two formations in different areas leads to the following conclusions.

a) The Kugitang formation. In spite of the variation in the strike of the beds, the deviation of the northwest and northeast trending joint systems is slight and does not exceed 14° to 15°. The angle between the two joint systems remains constant.

The submeridional and sublatitudinal joints (IIa and IIb) give maxima in the fourth area. The northwest trending joints (Ib) are much less strongly developed in the fourth than in the second area.

b) The Gaurdak formation. The orientation of the northwest and northeast trending joints (Ia and Ib) changes very slightly from one area to another (4° to 9°). The dips are somewhat steeper in the third area and the angles between the joint systems decrease from the second to the fourth area by only 10°.

The orientation of the joint systems of the submeridional and sublatitudinal trends (IIa and IIb) also changes only slightly, although a little more than in the preceding case, the maximum deviation being in the

Area	Formation	Attitude of beds			Joint systems											
					I						II					
		Azimuth		Angle of dip	a			b			a			b		
					Azimuth		angle of dip	Azimuth		angle of dip	Azimuth		angle of dip	Azimuth		angle of dip
		dip	strike		dip	strike		dip	strike		dip	strike		dip	strike	
Second	G	202	292	18	46	316	59	126	36	75	348	78	80	90	0	79
	K	206	296	20	41	311	81	116	26	80	Not found					
Fourth	G	274	4	14	50	320	78	120	30	70	343	73	82	89	359	70
	K	264	354	16	31	301	80	139	49	81	347	77	75	93	3	70

third area (11^{o} to 14^{o}), whereas the difference in strike in the second and fourth areas amounts to only 1^{o} to 3^{o}. The dips and the angles between the joint systems also change very little (5^{o} to 10^{o}).

The dips of the joints in the Kugitang and Gaurdak formations are either equal or are somewhat steeper in the former. The following data have been obtained on the relative age of the joint systems. The northwest trending joints cut the northeast trending joints; i.e., the joint system Ia is younger than Ib. Furthermore, the Ib joints are displaced by the submeridional and sublatitudinal joints (IIa), i.e., the former are older. The relationships among the other systems remain undetermined. The relation of the joint system trends to the structure of the dome is shown in Table 1 and in Fig. 5.

Thus, within the Gaurdak sulfur deposit there are, besides the bedding joints, four joint systems: with northwest, northeast, submeridional and sublatitudinal trends. The trend of a joint system does not change either in passing from one formation into the other or with its position in the structure.

It follows from this that the jointing was superimposed on the already completed structure, for joints developed before or during the formation of the dome would have changed their orientation, depending on their position on its limbs; in other words, the joint systems are younger than the dome.

A comparison of the directions of the joints and of the two large faults in the immediate vicinity of the dome shows that of the four joint systems, two are parallel to the faults; this is well shown in Table 2.

It is noteworthy that the greatest maxima of the joint system Ia correspond to the largest displacement of the fault striking northwest. The submeridional joints may be correlated with the similar Kugitang faults. The relatively weak maximum IIb may be the result of the considerable distance (35 to 45 km) of the Gaurdak dome from these faults.

The origin of the sublatitudinal joints which give the IIa maximum and their relation to the tectonics of the region are not known.

On the basis of these facts we may attempt to determine the time of formation of the joint systems and of the Gaurdak dome. We shall assume that the faults and the joint systems, at least those that give

Table 2

Fault	Dis-placement	Strike of fault	Strike of joints						Joint system
			2nd area		3rd area		4th area		
			Gaurdak fm.	Kugitang fm.	Gaurdak fm.	Kugitang fm.	Gaurdak fm.	Kugitang fm.	
Uzun-Kuduk graben (Normal fault and thrust?)	1500-2000	315°	316	311	311		320	301	Ia
Hinge fault	500-800	40°	36	26	28		30	40	Ib

the Ia, Ib and IIb maxima, were formed at the same time.

The upwarping evidently began at the end of the Lusitanian and continued into the Kimmeridgian and Tithonian as a part of the neo-Kimmeridgian phase of folding. This is reflected in the presence of an unconformity between the Kugitang and Gaurdak formations and, especially, in the thinning out of the Gaurdak formation from the periphery to the center of the dome. The possibility of such thinning out by diapiric extrusion or solution of the anhydrites cannot be entertained in view of the attitude of the key bed. The angle between it and the roof of the Kugitang limestones decreases towards the center of the dome and disturbances in its attitude because of solution of the underlying rocks are rare and local (Fig. 2). During the Cretaceous and Paleogene, the upwarping ceased and the Gaurdak dome participated in the general subsidence of the region. The inception of the now existing dome must have occurred at the end of the Paleogene or the beginning of the Miocene as an episode in the general uplift of the Gaurdak-Kugitang region. Judging by the dislocations in the early Quaternary Kan-Say conglomerates, this uplift continued into the Quaternary. The fact that the remnants of fanglomerates on the watersheds between the canyons of the Gaurdak dome lie 500 to 600 m above their base in the Kan-Say Valley and that farther to the southwest they are found in the lower ends of these canyons indicates that at the time of their accumulation the uplift had already reached considerable height (a minimum of 2,000 m of Cretaceous and Paleogene rocks had been eroded.). The trend of the valleys has only an indistinct relationship with the trend of jointing, or perhaps none at all, indicating that the existing valleys of the Gaurdak were initiated before the formation of the joints. It may be assumed that the dome came into existence in Neogene time toward the beginning of the Quaternary and that the faults and joints formed later. This does not exclude N.P. Petrov's view [5] on the fluctuations of the erosion level during the Quaternary and the resulting periods of erosion and deposition. The relatively small movements with amplitudes of a few hundred meters which occurred after the formation of joints did not seriously affect their trends. It is possible the development of cleavage along the northwest trending joints is the result of these movements.

It was mentioned above that many geologists believe that fractures play an important role in the formation of sulfur deposits in general and those of the Gaurdak dome in particular. We shall attempt to determine whether there is a connection between the distribution of the ore bodies and the dislocation tectonics at the Gaurdak deposit. Even a cursory examination of the outlines of the ore bodies (Fig. 6) shows that their trends follow a system. Comparison of Figs. 5 and 6 shows that the trends of the ore bodies do not depend directly on the structure of the dome. It may be supposed that they depend on the direction of the faults. During a recent exploration in the third area a fault striking 290° to 300° was found paralleling the elongation of the ore body. Along this fault the thickness of the sulfur-bearing beds and the content of sulfur are at a maximum. The fact that this fault was discovered only after many years of exploratory work confirms what has been said earlier about the difficulty of discovering faults in the Gaurdak formation. It was shown above that at least some of the joints have developed parallel to the faults. If the ore bodies are actually related to the faults, there must be a relation

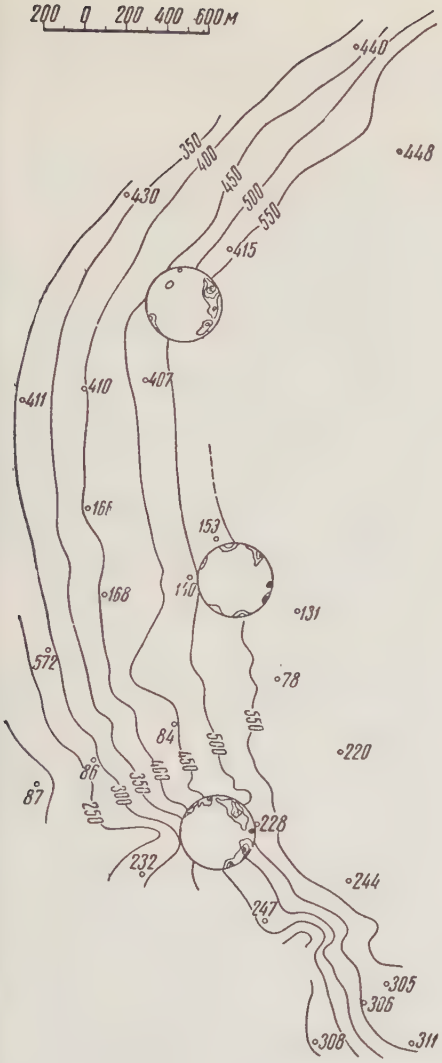

FIGURE 5. Structure of the top of the key bed (R). Circles are joint diagrams for the Gaurdak formation.

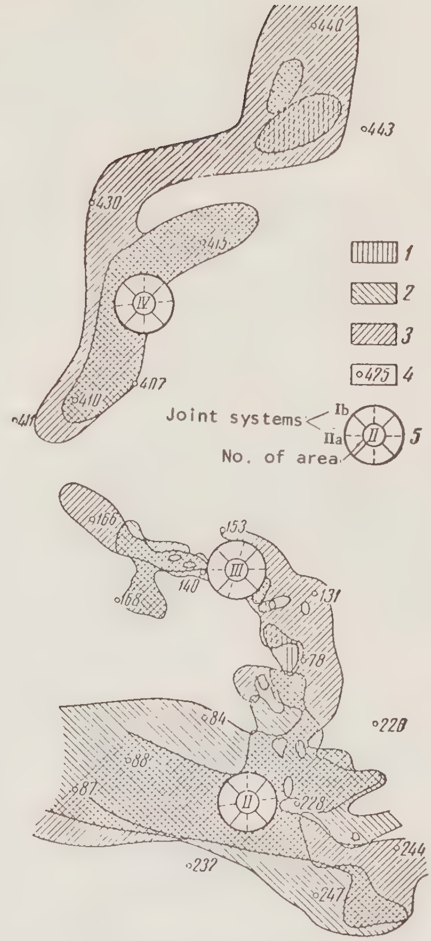

FIGURE 6. Outlines of the ore bodies.

1 -- Bed R; 2 -- Bed D; 3 -- Bed F; 4 -- Borehole; 5 -- Joint system and number of area.

between the trends of the ore bodies and the joints. That this relation exists has been confirmed at Gaurdak. The results of correlation of the trends of the joints and the ore bodies are given in Table 3.

Our data show that there are two trends among the ore bodies in different areas, the sublatitudinal and the submeridional, which agree well with the trends of the corresponding joint systems (IIa and IIb). Besides this, there are ore bodies with northwest trend (second and fourth areas) and northeast trend (fourth area) correlatable with the joint systems Ia and Ib. The difference between the trends of the ore bodies and the joints varies from 2° to 16°. An interesting detail should be mentioned. The difference in the trends of the ore bodies and the corresponding joint systems is less in the Kugitang than in the Gaurdak formation (see Table 1). If we take this into consideration, this difference varies from 2° to 11° and on the average amounts to 7° as against 9° if the jointing of the Kugitang formation is neglected. The slight discrepancy in the trends of the ore bodies and joints may be the result to some extent, of errors of measurement, or, as was shown by S. V. Nikolayev [4] in connection with the carbonate rocks of the Samarskaya Luka region, to the change in

Table 3

Trend of the ore bodies, degrees			Second area			Third area		Fourth area		
			85	300	355	315	0	75	45	10
Joint trends, degrees	I	a		316		311			30	
		b			36		38			320
	II	a	78			88		73		
		b			0		14			359
Difference between joint trends and trends of ore bodies, degrees	I	a		16		4			15	
		b								
	II	a	7					2		
		b			5		14			11

azimuth of the joints in passing through strata with different structures. This should be true also of beds with different lithologies.

This evidently explains also the slight discrepancy in the azimuths of the same joint systems in the Kugitang and the Gaurdak formations.

The comparison of trends of the ore bodies and the joints reveals a number of other interesting features. First, the ore bodies tend to be elongated parallel to the sublatitudinal and submeridional directions; i.e., they are most clearly related to the joint systems which give the least pronounced maxima (IIa and IIb). Second, in the second area, where the trend of one of the ore bodies coincides with the northwest trending joint system, there is a slight discrepancy between the trend of the ore body and the main maximum of the system. In this case the trend of the ore body coincides with the secondary maximum; i.e., again there is a tendency of the ore bodies to parallel the less strongly developed joint systems. Interesting also is the absence of northwest trending ore bodies in the fourth area. It is hoped that further investigation of the age relationships and morphology of the joints will solve this and other problems.

Our data show that there is an obvious relation between the trends of the ore bodies and joints in the Gaurdak sulfur deposit.

However, this may be the result of a number of causes. It may be because of the concentration of sulfur previously dispersed over the large area where it was formed syngenetically, or it may be because of secondary formation of sulfur along definite tectonic elements. If the first hypothesis is true, dispersed syngenetic sulfur must exist in the limestones beyond the boundaries of the deposit. Actually these limestones are very dense; the anhydrite enclosed in them does not become hydrated, and like the limestones, it is devoid of free sulfur. Locally, especially in the southern part of the deposit (first area), there are relicts of a former accumulation of sulfur. The structure and other features of these relicts suggest that the sulfur which migrated away was definitely epigenetic, for it was associated with caverns and breccias in the limestones. These relicts fit well into the general plan of distribution of sulfur in the present deposit. The migration of sulfur occurs here in the form of SO_2 which is

immediately oxidized into SO_3. The oxidation of sulfur in this part of the deposit is the result of unfavorable geochemical conditions.

Thus, it is more probable that sulfur is of epigenetic origin. It should be noted that the problem of the genesis of sulfur, as of any other economic mineral, cannot be reduced to the elucidation of fracture tectonics of the locality, especially if this is represented only by joints. It is evident that the problem or origin can be solved only on the basis of data on tectonics, mineralogy, etc. Therefore, the data given here do not provide a final solution of the problem of the origin of the deposit but point only to the greater likelihood of epigenetic origin of sulfur. They lead to the conclusion that the accumulation of sulfur occurred after the formation of the dome and of regional faults and joint systems, i.e., not earlier than the Pleistocene.

In summary, our observations:

1) provide new data which confirm an earlier view that the northwest fault antedates the Gaurdak dome [2, 6],

2) establish an intimate relationship between deposition of sulfur and certain joint systems,

3) confirm the epigenetic origin of the Gaurdak sulfur deposit and its Quaternary age,

4) give reasons to believe that studies of fracture tectonics by the methods of B.P. Belikov [1] and A. Ye. Mikhaylov [3] are useful in the very first stages of investigation of sulfur deposits. However, in every case the individual structural features of a given area must be taken into consideration.

The author expresses his deep appreciation to B.P. Belikov and Yu. A. Rozanov for their advice and criticism during the reading of the manuscript.

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BRIEF COMMUNICATIONS

ON THE DETERMINATION OF THE ABSOLUTE AGE OF POTASH FELDSPARS BY THE ARGON METHOD

by Kh. I. Amirkhanov, S.B. Brandt, and Ye. N. Bartnitskiy

It was discovered by E.K. Gerling (1952) that radiogenic argon escapes more readily from feldspars than from micas. The curves showing loss of argon as a function of temperature indicate that feldspars begin to lose argon at low temperatures. Soviet experience in geochronology has confirmed these facts. Recently, scientists abroad also have come to this conclusion [2, 4].

The low-temperature loss of argon from feldspars is explained by their tendency to alteration. Centers of sericitization and kaolinization may appear anywhere within a feldspar crystal resulting in the formation of mica and kaolinite. The formation of perthite may be also regarded as a process of alteration. All processes which result in a reconstruction of the crystal lattice may cause loss of radiogenic argon.

The loss may be explained in a number of ways. For example, W. Gentner and W. Kley [2] consider that the "holes" which form as a result of these processes and the boundary surfaces of the altered zones serve as loci of accumulation of radiogenic argon which escapes from the crystals by volume diffusion. A proof of this is seen in the fact that very fine fragmentation which takes place mainly along the altered surfaces is accompanied by spontaneous loss of argon.

This explanation seems very unlikely. The diffusion constant of the micas is very small (10^{-3} cm^2/sec) and there is no reason to believe that it is greater in the single crystals of feldspar [1]. Besides, if the concentration of radiogenic argon in the centers of accumulation were to exceed its concentration in the crystal as a whole, which must happen sooner or later, diffusion would stop. Therefore, using micas as an illustration, it may be assumed by analogy that the low-temperature loss of argon from the feldspars results also from the escape of argon atoms lying at the boundaries between individual crystals. Therefore, the cause of argon loss is increase in the specific area of the feldspar. It may be also supposed that the altered regions, because of various reconstructions of the feldspar lattice, become "transparent" to the radiogenic argon.

In spite of these shortcomings the use of feldspars for age determination has considerable possibilities. Hence, the great amount of effort devoted by scientists to the development of criteria for the degree of preservation of argon in feldspars.

If all altered areas in a sample of feldspar in which the radiogenic argon is unstable is united into one unstable zone, and the areas in which the structure has not been altered and from which the escape of argon is improbable are also united into a single zone, then it may be assumed that the indefiniteness in the age determination is the result of potassium and argon contained in the unstable zone. If, therefore, the volumes of the stable and unstable zones in the sample were known before the age determination, a criterion of the stability of radiogenic argon would be found. Evidently the $A^{40} : K^{40}$ ratio in the stable zone of a feldspar would give the true age of the sample. However, at present there are no methods by which the two zones can be separated.

Let us consider the possibility of complete removal of radiogenic argon and potassium from the unstable zone. Then in our measurements we shall deal only with A^{40} and K^{40} of the stable zone. The removal of A^{40} from the unstable zone may be accomplished by heating the sample to a low temperature. As for the removal of potassium, it can be accomplished by chemical treatment of the powdered sample in which the altered areas will be sufficiently well exposed.

We have experimented with this method on a sample of Precambrian feldspar from Karelia and on a sample of phlogopite. Treatment of the feldspar samples with hydrochloric acid of various concentrations and with a solution of potassium chloride under normal and elevated temperatures and pressures had no effect on the potassium content. Therefore, it was decided to treat the samples with thallium nitrate under both normal and elevated temperatures and pressures. Gruner

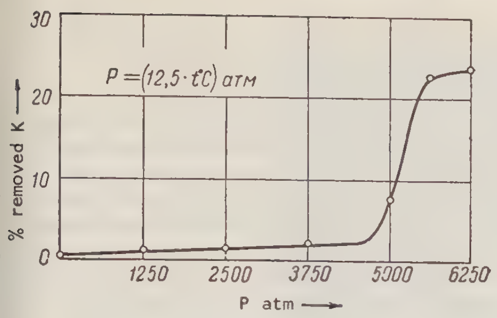

FIGURE 1.

sed this treatment on glauconites, but only nder normal pressure [3].

In order to obtain high pressures, a steel omb was constructed with working volume of he order of 10 cm^3. A definite quantity usually 1.0 g) of the sample was placed in he bomb with the amount of thallium nitrate eeded to obtain a saturated solution, and the omb was filled with water. The "loaded" omb was heated for three hours at a definite emperature, and the thermal expansion of ne solution in the bomb produced the required ressure. After this treatment, the amount f potassium removed from the sample was etermined.

The data obtained from the 0.03 to 0.05 nm fraction are given in Figure 1, where ne amount of removed potassium is shown s a function of pressure. The pressure was alculated approximately by the formula $\approx 12.5 \times t^{o}$ C atm., based on the thermal xpansion of water in the bomb. It will be een that at pressures corresponding to 450° nd more the amount of replaced potassium eached 23 percent and remained constant rith further increase in pressure. For the raction with grain size <0.03 mm, the mount of removed potassium at t = 450° was bout 32 percent.

In order to determine the effect of temerature on the degree of replacement of otassium, the experiments were repeated rith only approximately half of the bomb eing filled. In this case no passage of potassium into the solution occurred up to t = 500°. Thus it is quite obvious that the replacement of potassium by thallium depends entirely on pressure and not on temperature.

Experiments were also performed with phlogopite (sp. 250/53, Yakutsk District, A.S.S.R., Aldan region, M.A. Liparev). The 0.03 to 0.05 mm fraction was treated with a solution of thallium nitrate for 3 hours at 500°. This resulted in the replacement of 13 percent of potassium. The amount of argon present in the "unstable zone" at this temperature is 16 percent [1].

It is important to note that the absolute ages determined from the initial content of argon and potassium

$$\frac{A^{40}\ \frac{n\ mm^3}{g}}{K\%} = \frac{1.070}{8.7} = 0.123$$

and from the remaining content

$$\frac{A^{40}\ \frac{n\ mm^3}{g}}{K\%} = \frac{0.900}{7.6} = 0.119$$

are almost the same. This confirms once more the fact that the low-temperature loss of argon in micas is due to desorption, i.e., a certain amount of potassium, and therefore of argon, is present in the surface layer of the mineral.

In conclusion we shall cite the absolute ages for a feldspar sample from Karelia computed on the basis of the total and "corrected" potassium content. (See Table 1.)

Although this research is not yet completed, it may be hoped that the method offered here, besides giving the absolute age of feldspars, will also provide some information on their structure. The method may be used with many other minerals and rocks.

Table 1

$TlNO_3$ treatment	$A^{40}\ \frac{n\ mm^3}{g}$	K%	Age in millions of years	Fraction, mm
Before treatment	0.892	10.7	1375	—
After treatment	0.892	8.2	1650	0.03 ÷ 0.05
After treatment	0.892	7.2	1800	<0.03

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ACCESSORY ORTHOLITE FROM ACTINOLITE ROCKS OF THE MALAYA LABA RIVER

by V.V. Ploshko

In 1956, the author made a detailed petrographic study of an area with an interesting mineral association on the left bank of the Malaya Laba River in the region of the Peredovoy Range (Northern Caucasus). The area was discovered by A.V. Netreba and described by G.D. Afanas'yev [2].

At the contact between the granites of the Urushtensk magmatic complex and albitized actinolite rocks of the same complex, metasomatic veinlets as much as 2 cm wide were found in which a mineral of the orthite group is associated with apatite, zircon, albite, and a rare earth phosphate (γ = 1.837, α = 1.787, $\gamma - \alpha$ = 0.050; negative; 2V very small; G = 4.683). The mineral of the orthite group occurs in the albitized actinolite rocks. Its properties are very similar to those of the orthite-like mineral from the carbonate veins of the same area which were described by Afanas'yev and to the orthite from the Magisho Range [2]. The physical and chemical properties of the mineral indicate that it belongs to the orthite group, but it has a number of characteristics which distinguish it from the typical members of the group.

The orthite occurs in granular aggregates of small prismatic crystals 6 mm in length. Some grains of the mineral are dark-green and translucent on the edges and others are light-green and completely transparent in small fragments; the luster is vitreous. The mineral is brittle, has an irregular fracture and a green streak. Small apatite crystals occur in it as inclusions (Fig. 1).

The X-ray diffraction data obtained by M.T. Yanchenko are given in Table 1. Comparison of the powder photograph data of our mineral with the standard data on orthite [4] shows similarity but not identity of structure. The main difference is in the variation of intensity of reflections, absence of some weak reflections present in the standard, and the presence of additional reflections (d = 1.247, d = 1.303, d = 1.420, d = 1.465, d = 2.43).

In thin sections the orthite is brownish-

Table 1

No.	I	21 var.	d	No.	I	21 var.	d
0	6	20,0	4,57	8	8	43,0	2,13
1	4	26,0	3,50	9	7	49,0	1,878
2	2	28,2	3,25	10	9	56,8	1,634
3	10	31,2	2,92	11	3	64,0	1,465
4	9	33,9	2,68	12	3	66,2	1,420
5	4	34,8	2,62	13	1	73,0	1,303
6	1	37,5	2,43	14	1	74,9	1,274
7	1	39,0	2,34				

Note: Comma represents decimal point.

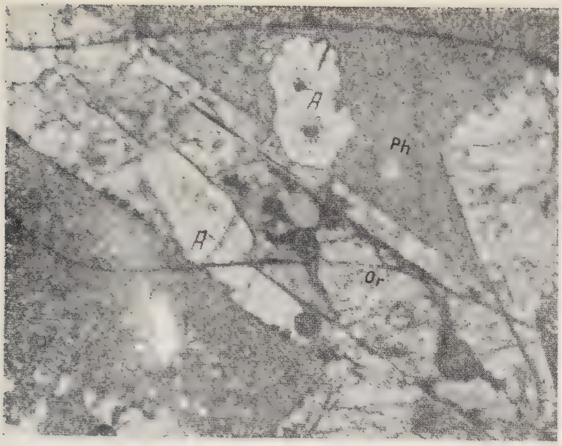

FIGURE 1. Orthite from metasomatic veinlets in actinolite rocks

Without the analyzer. 46X, (Or - orthite; Ph - rare earth phosphate; A - apatite).

green or greenish, sometimes colorless. The distribution of color in the grains is irregular, spotty, occasionally zonal. Pleochroism is strong. Zoning is commonly emphasized by the presence of the products of metamictization. There are grains which are completely altered into a brown isotropic mass.

The mineral is biaxial, positive, less commonly negative; monoclinic (parallel extinction on the (001) cleavage traces, $\underline{X} \wedge \underline{c}$ = 27 - 33°). The (110) cleavage is more distinct than the (010). Dispersion of the optic axes distinct (p > v). Elongation is either positive or negative, since $\underline{X}$ is parallel to the length of the grains.

Table 2 presents the physical properties of orthite from the actinolite rocks and carbonate veins of the Malaya Laba River, from the Magisho Range (Northern Caucasus), and from Bektau-Ata Mountain in Kazakhstan. An examination of the table shows that the physical properties of the orthites from Northern Caucasus deviate somewhat from those of orthite and trend towards clinozoisite.

When powdered, orthite is slightly gelatinized by hot concentrated hydrochloric acid, indicating a degree of metamictization [6].

A complete chemical analysis of orthite from the actinolite rocks of the Malaya Laba River was made by L. B. Tumilovich in the laboratory of the IGEM Academy of Sciences U. S. S. R. (Table 3). Analyses of orthites from the carbonate veins of the same locality, from the Magisho Range and from Bektau-Ata Mountain are given for comparison.

The table shows that the orthites from the Malaya Laba River have some elements which do not occur in the minerals of the orthite group, namely fluorine and phosphorus. This is evidently, because, in spite of careful selection of material under a binocular magnifier, it was not completely free from minute inclusions of apatite. For this reason, F and P, together with the amount of Ca required by the apatite formula, were excluded from the analysis, and the analysis was recalculated to 100 percent.

The table shows that the orthites of Northern Caucasus are very similar chemically and that their distinctive feature and main difference from typical orthite is a certain deviation towards clinozoisite. The orthite from the actinolite rocks differs from the Bektau Ata orthite by its small MnO content and relatively high MgO content. Unlike the Bektau-Ata orthite, which contains little water, our orthite has a high water content. In this respect it approaches the orthite from the Rubidoux Mountains, California [7]. It is characterized also by a high ThO_2 content, and in this resembles the orthite from Ryozen, Japan [6]. Its rather high water content is due, probably, to the presence of thorium. Many mineralogists who have studied metamict orthites [7] believe that the alteration is due to uranium and thorium radiation, which breaks ionic bonds and favors penetration of water into the crystal lattice.

The Malaya Laba River orthite contains a noticeable amount of Na_2O, which also points to the change in its original composition, for the presence of alkalies is usually characteristic of metamict orthites [5]. The presence of SrO is a general feature of the orthites from Northern Caucasus and is

Table 2

Optical Constant	Orthite from actinolite rocks (Malaya Laba River, N. Caucasus)	Orthite from carbonate veins (Maya Laba R., N. Caucasus)	Orthite from the Magisho Range (N. Caucasus)	Bektau-Ata orthite (Kazakhstan)[4]
2V	± 70-90°, commonly 80-86°	Positive	Positive	Negative
Refractive indices				
γ	(1.713-1.718) ± 0.001	1.740[1]	1.680[2]	1.758
β	(1.709-1.714) ± 0.002	—	—	—
α	(1.700-1.704) ± 0.001	1.728	1.665[2]	1.724
$\gamma - \alpha$	0.006-0.018	0.012	0.015	0.034
Pleochroism				
γ	Yellow, grayish-yellow, greenish-brown	Greenish-brown	Grayish-yellow with a greenish tinge	Dark brown red
β	Light-yellow with greenish cast, grayish-green	Grayish-green	Faintly greenish	—
α	Light-yellow, almost colorless, yellow with greenish cast	Yellow with greenish cast	Light-yellow, almost colorless	Light olive green to almost colorless
Absorption	$\gamma > \beta > \alpha$	—	—	—
Specific gravity	3.496[3]	—	—	4.16

[1]G. D. Afanas'yev [2].
[2]G. D. Afanas'yev [1].
[3]Determined by V. S. Amelina.
[4]Ye. Ye. Kostyleva and M. C. Kazakova [3].

characteristic in general of the epidote-zoisite group of minerals in the rocks of the Urushten magmatic complex [1].

In general, chemical analyses fit the formula of orthite: (RE, Ca)$_2$ (Al, Fe)$_3$Si$_3$O$_{12}$ (OH). The somewhat low total of Ca and rare earth oxides and the high water content are probably the result of the penetration of water into the crystal lattice during metamictization. Moreover, it should be borne in mind that analyses of metamict orthites commonly show an excess of SiO_2 over that present in the structure of the mineral [5], and for this reason there is always an apparent deficiency of Ca and RE cations as compared with silicon. It is possible also that this deficiency may be the result of a structural peculiarity of the mineral which distinguishes it from the typical orthite.

The rare earths separated in the course of chemical analysis were identified in the X-ray and Spectrographic Laboratory, IGEM Academy of Sciences U. S. S. R., by L. A. Voronova. The following percentages of the rare earths were found in the orthite: (CeO_3) 42.9, (La_2O_3) 29.4, (Nd_2O_3) 15.3, Pr_2O_3) 11.0, (Sm_2O_3) 1.2, (Gd_2O_3) 1.0 and (Dy_2O_3) 0.1

Besides the elements determined chemically, the orthite contains Be, Ga, V, Cu, Ni, Zr, and Ba, present in thousandths of one percent and determined spectrographically by P.I. Sumina.

Thus the orthite from Northern Caucasus described here in detail for the first time differs in its physical properties and chemical composition from the typical orthites and approaches clinozoisite. Especially interesting is its association with the rare-earth phosphate of late crystallization. The Malaya Laba River orthite is most commonly found in the contact zones between the actinolite rocks and granites of the Urushten magmatic complex, which are characterized by the presence of accessory minerals of the epidote-orthite group. As is shown by the experience of many investigators, orthites are usually associated with granitic magma and its derivates and are most frequently

Table 3

Oxides	Orthite from actinolites of the Malaya Laba River (Northern Caucasus)			Orthite with apatite inclusions from the carbonate veins, Malaya Laba River	Orthite from the Magisho Range (North Caucasus)	Orthite from Bektau-Ata Mountain (Kazakhstan)
	% before exclusion of apatite	% after exclusion of apatite	Atomic ratios	%	%	%
SiO_2	30,96	32,85	0,5469	26,50	35,04	30,47
TiO_2	0,20	0,21	0,0026	0,22	0,10	1,39
Al_2O_3	18,24	19,36	0,3798	12,78	23,22	12,43
Fe_2O_3	5,64	5,98	0,0748	6,23	4,18	7,29
FeO	4,29	4,55	0,0637	4,05	4,52	10,93
MgO	2,45	2,60	0,0644	2,51	0,13	0,50
MnO	0,08	0,09	0,0012	0,04	0,21	1,88
CaO	11,86	9,67	0,1724	19,93	17,60	9,69
BaO	—	—	—	—	Not determined	—
SrO	0,52	0,55	0,0053	0,54	0,42	—
Na_2O	0,33	0,35	0,0114	0,28	—	—
K_2O	—	—	—	none	—	—
ΣTR_2O_3	17,08	1,813	0,1098	11,48	14,20	—
Ce_2O_3	—	—	—	—	—	12,70
$(La, Dy)_2O_3$	—	—	—	—	—	12,06
Y_2O_3	—	—	—	—	—	0,22
ThO_2	2,28	2,42	0,0092	1,16	—	—
H_2O^-	0,44	0,46	—	0,49	—	0,10
H_2O^+	2,62	2,78	0,3086	2,36	—	0,29
P_2O_5	2,10	—	—	9,07	Not determined	—
F	0,18	—	—	0,78	—	—
CO_2	—	—	—	2,02	—	—
Total	99,27 —0,07F 99,20	100,00		100,44 —0,32F 100,12	—	99,95
O	—	—	2,4163	—	—	—
Analyst	A. B. Tumilovich (1958)			A. B. Tumilovich (1958)	A. B. Tumilovich [2]	M. E. Kazakova [3]
Reference					G. D. Afanas'yev [2]	Ye. Ye. Kostyleva and M. Ye. Kazakova [3]

Note: Comma represents decimal point.

und in pegmatites; hydrothermal orthites re very rare [8]. The orthitelike material escribed by Afanas'yev from the carbonate eins is undoubtedly hydrothermal [2]. Evidently, the orthite from the metasomatic einlets in actinolite rocks is analogous to e orthite from the carbonate veins of the ame area. In any case, there is no doubt f its connection with the metasomatic effect f granitic magma of the Urushten magmatic omplex on the enclosing basic rocks.

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STRUCTURAL CHARACTERISTICS OF THE SOUTHERN LIMB OF THE BELOMECHET SYNCLINE IN NORTHERN CAUCASUS

by R. S. Bezborodov

The Belomechet syncline, the eastern part of the East-Kuban' downwarp [5], stands out clearly on geologic maps as an area of Neogene and Quaternary sediments in the midst of Paleogene deposits.

In recent years the Kavkaznefterazvedka Organization drilled a number of deep exploratory holes on the southern limb of the Belmochet syncline in the Frolovskiy and Cherkessk areas (Fig. 1).

An analysis of cores and electric logs from the boreholes in these areas and the correlation of Lower and Middle Jurassic sections within these areas with sections studied by the author in the nearby uplands, make possible a number of conclusions about the structural characteristics of the southern limb of the Belomechet syncline and the time of formation of the downfold.

The Lower and Middle Jurassic deposits exposed on the northern slope of the Caucasus between the Kuban' and Bol'shoy Zelenchuk Rivers may be divided into three large stratigraphic-lithological complexes.

Middle Liassic (Pliensbachian stage). The Jurassic section of the region begins with a series of sandstones about 700 m in thickness which are referred to the Pliensbachian stage on the basis of G. Ye. Pilyuchenko's data [4]. The series consists of groups of beds and single beds of medium- and coarse-grained, locally gravelly sandstones of various thicknesses and dark-gray, almost black argillites.

On the Kuban' River the lower part of the section contains andesitic sills.

Here the Pliensbachian rocks are coal-bearing (Khumara coal deposits).

Upper Liassic (Toarcian and lower Aalenian stages). The Pliensbachian beds in the Kuban' River basin (in the southern part of the region) are unconformably overlain by volcanic rocks of the lower Toarcian age [2]; these rocks are found only locally and are absent to the north of the village of Khumara.

Higher in the section, resting on strata of different ages, lies an onlap series of poorly sorted, sandy and silty, shallow-water rocks of "debris-like" aspect containing layers of conglomerate. The lower part of this series contains upper Toarcian ammonites and the upper part, lower Aalenian ammonites. The thickness of the upper Toarcian-lower Aalenian beds on the Bol'shoy Zelenchuk River is 150 to 160 m and on the Kuban' River, 40 to 50 m. In the latter locality these beds decrease rapidly in thickness to the north, and in the vicinity of the village of Khumara, wedge out of the section.

Middle Jurassic (upper Aalenian substage and the Bajocian stage). The upper Aalenian sediments have been identified by their fauna only in the Kuban' River basin, where they form a 15 to 20 m sequence of gray argillites truncated in the north by the transgressive Bajocian deposits.

In the Bol'shoy Zelenchuk River section,

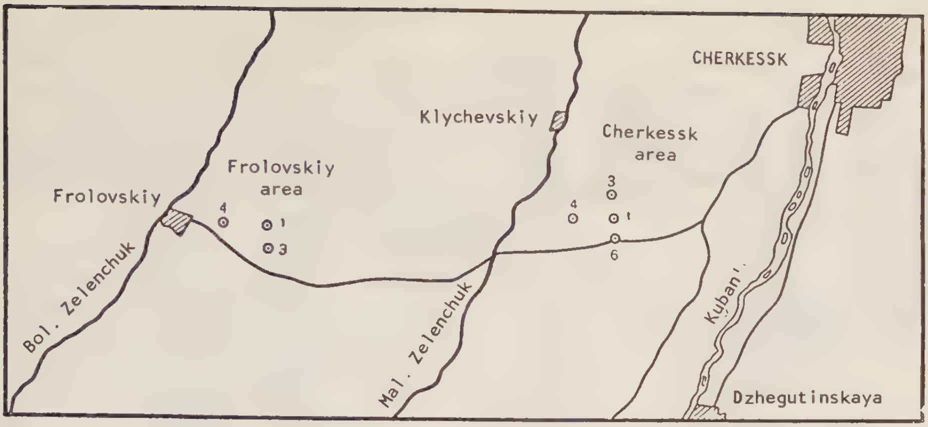

FIGURE 1. Distribution of boreholes in the Cherkessk and Frolovskiy areas

he upper Aalenian deposits are evidently ıbsent, because very near the base of the Bajocian sediments the author found numer- ous lower Aalenian Leioceras opalinum Rein.

Higher in the section lies a thin group of crinoidal and arenaceous limestones which pass into conglomerate in the north. According to I. R. Kakhadze and others [2], this group is of Bajocian age. Finally, the upper part of the Middle Jurassic section is represented by gray and bluish-gray argillites with concretions of argillaceous siderite and a Bajocian fauna (mainly upper Bajocian) [1, 2]. The thickness of the Bajocian sediments on the Kuban' River is 400 to 500 m and on the Bol'shoy Zelenchuk River, 600 to 700 m.

The Bajocian argillites are overlain by Upper Jurassic limestones and redbeds with a thin sequence of sandy and gravelly Callovian beds at the base.

In the Frolovskiy area, borehold No. 1, after passing through Cretaceous and Upper Jurassic beds, in the 1,250 to 1,530 m interval[1] penetrated through a sequence of gray argillites with argillaceous sideritic concretions of typical Middle Jurassic aspect. At the depth of 1,420 to 1,440 m, the argillites contain a conglomeratic zone 10 to 20 m in thickness (Fig. 2).

Beneath the argillaceous Middle Jurassic sediments at the depth of 1,530 to 1,730 m, lies a series of sands, silts, and clays with irregular lenticular and torrential cross-bedded stratification. The rocks are commonly micaceous and contain numerous fragments of carbonized wood. Some beds contain gravel. Lithologically the rocks in this part of the section are like the Toarcian-lower Aalenian beds which outcrop farther south in the uplands and are tentatively referred by the author to the upper Liassic. Examination of cores from borehole No. 1 shows that the dips in the entire Upper Liassic and Middle Jurassic section do not exceed 5^{o} to 10^{o}.

Lower in the section, the borehold passed through a sequence of quartz conglomerates, quartz-feldspar-mica, coarse-grained and medium-grained sandstones containing layers and groups of beds of gravelly, almost black argillites with inclusions of carbonized plant debris. The entire sequence is enriched in small pebbles and granules and in boulders of coarse-grained pink and grayish-pink granite (of the Malkinskiy type). The number of granite pebbles and boulders increases towards the base of the sequence. Borehole No. 1 passed through 852 m of this sandy-gravelly material and was stopped in it at the depth of 2,582 m.

The study of the cores showed that the rocks in this part of the section (argillites in particular) are more compacted and metamorphosed than the overlying sediments. Moreover, the rocks of this sequence dip at 20^{o} to 30^{o}, i.e., considerably more steeply than the younger rocks. All this, together with the littoral, coal-bearing character of the sediments, permits us to refer them to the middle Liassic and regard them as analogues of the Pliensbachian coal-bearing beds outcropping in the more southerly, especially, southeasterly regions.

Borehole No. 4 (Fig. 2), 1.8 to 1.9 km to the west of borehole No. 1, has a similar section (Fig. 1).

[1]Depth from lip of the hole; relief is disregarded.

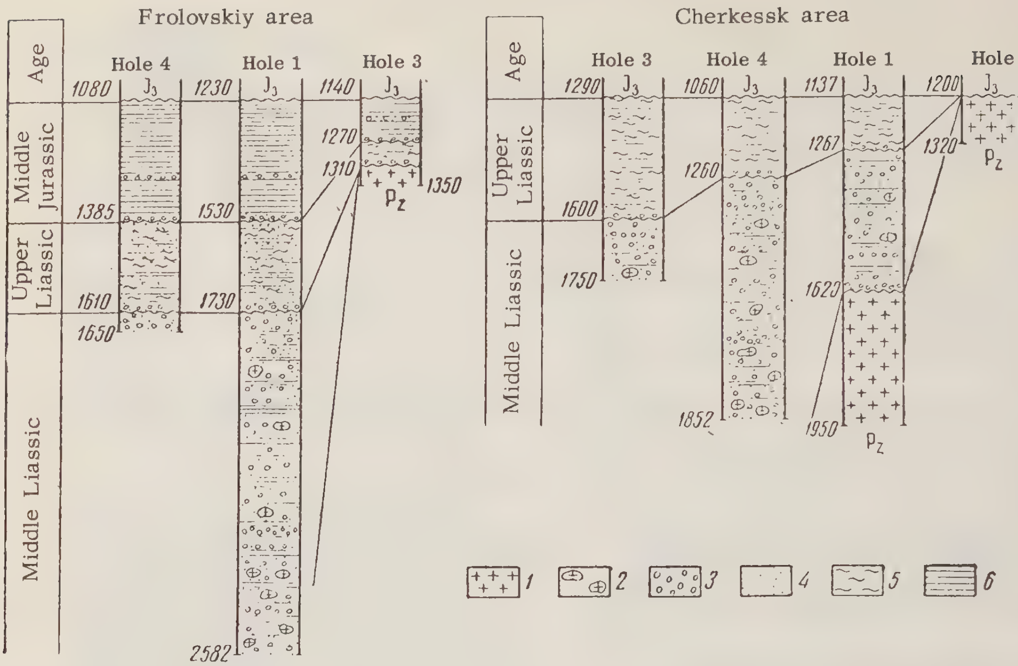

FIGURE 2. Correlation of borehole sections

1 -- granite; 2 -- granite pebbles and boulders; 3 -- pebble conglomerate; 4 -- sandstone; 5 -- siltstone; 6 -- argillite

Borehole No. 3, 0.9 km south of borehole No. 1, is entirely different. After passing through Upper Jurassic limestones, in the 1,140 to 1,270 m interval, it penetrated through a group of Middle Jurassic gray argillites with argillaceous sideritic concretions and a sandy-gravelly layer lying approximately in the middle of the section. Farther, at the depth of 1,270 to 1,310 m, the section is represented by a group of beds completely analogous to those encountered in borehole No. 1 at the depth of 1,530 to 1,730 m and and tentatively referred by the author to the upper Liassic. At the depth of 1,310 m, borehole No. 3 entered into massive pink coarse-grained granites of the Malkinskiy type, which it penetrated for 50 m before being stopped (Fig. 2).

Thus, within the Frolovskiy area the basement rocks plunge deeply to the north, undoubtedly because of the presence of a fault on which the northern part of the region (the Belomechetsk syncline) underwent considerable subsidence in the Lower and Middle Jurassic times (Fig. 3). The increase in thickness of the Lower and Middle Jurassic strata towards the north between boreholes No. 3 and No. 1 amounts to only 0.9 km over the distance of 1.2 km.

No less interesting are the data obtained in drilling the Cherkassk area located farther east. Here, borehole No. 1, after emerging from the Upper Jurassic deposits at the depth of 1,137 m, did not encounter the characteristic Bajocian gray argillites as in the Frolovskiy area, but entered directly into rocks quite analogous lithologically to the shallow water sandy-silty formations, which were in the Frolovskiy area tentatively referred to the upper Liassic. The thickness of this presumably upper Liassic series penetrated by the borehole in the Cherkessk area is about 130 m. At the depth of 1,267 m, borehole No. 1 passed through a 350-meter sand-gravel sequence with pebbles and boulders of pink granite analogous to the middle Liassic series of the Frolovskiy area. Finally, at the depth of 1,620 m, the borehole entered monolithic pink granites and continued through them for 330 m (the bottom of the hole is at 1,950 m).

Beneath the Upper Jurassic deposits, borehole No. 3, drilled about 900 m north of borehole No. 1, disclosed 310 m of upper Liassic sandy and silty rocks and was stopped in the sandy and gravelly middle Liassic beds (140 m). Approximately the same section was obtained from borehole No. 4, located 1.7 to 1.8 km almost due west of

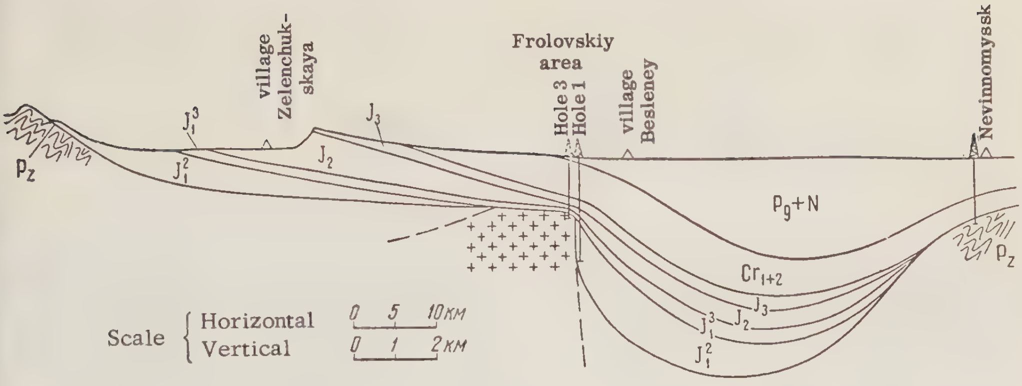

FIGURE 3. Section of the Frolovskiy area along the Bol'shoy Zelenchuk River.

borehole No. 1 (Fig. 2).

Thus, the steep southern limb of the Belomechet syncline is bounded by a fault on which movements continued during a rather long geologic time. Their intensity diminished gradually from the middle Liassic to the Upper Jurassic (Fig. 3).

The buried granitic range of sublatitudinal trend, evidently continues farther to the east as the Malkinsk-Mineralovodskoye intrusive in the region of the uplifted basement.

To the west, this zone appears as the Spokoynenskoye uplift on the Laba River. Still farther west, it is apparently reflected in a series of uplifts such as Yaroslavskoye, Kalininskoye, Belorechenskoye, etc., located on the northern and northeastern plunges of the Adygey salient.

The increase in thickness of the Lower and Middle Jurassic rocks of the modern Belomechet syncline indicates that this structure was first firmed in the Lower Jurassic and undoubtedly existed as a region of intensive downwarping during the middle Liassic time.

The North Caucasus downwarp was divided during the Liassic and Middle Jurassic time into two independent zones of subsidence with sublatitudinal trends. The southern zone corresponded to the modern northern Jurassic depression and the northern zone lay within the boundaries of the Belomechet downfold and of its western continuation. Both zones of downwarping were cut by a geanticlinal uplift of the Hercynian basement from which at certain times detrital material was carried both to the north and to the south.

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REVIEWS AND DISCUSSIONS

YE. S. DOBROKHOTOVA'S "PRACTICAL INSTRUCTIONS ON THE USE OF YE. S. FEDOROV'S METHOD IN PETROGRAPHY"[1]

by K.V. Ivanov

Although at present there are many manuals and texts on the Fedorov method (by A.K. Podnigin, Ye. A. Kuanetsov, V.S. Sobolev, G.M. Saranchina, N.A. Yeliseev, and others), there is still no adequate account of the techniques and methods actually used in work with the Fedorov stage. Saranchina's "Fedorov's Method" devoted the most attention to methods, but unfortunately, her book was published in a relatively small edition. Meanwhile the need for a manual on the Fedorov method which would present in the briefest possible form the basic techniques for determining plagioclases, alkali feldspars, and colored minerals grows yearly.

This need has recently been filled by Ye. S. Dobrokhotova's book, "Practical Instructions on the Use of Ye. S. Fedorov's 'Method in Petrography.'" In the introduction the author states, that the use of this manual during those few hours which are devoted to the universal stage in the curricula of the geological prospecting institutes, would shorten the time taken by instruction in techniques and allow more time for the methods of determining alkali feldspars and colored minerals.

The first chapter gives a brief but sufficiently complete account of the history of the Fedorov stage, its mounting and adjustment. A few points in this chapter require comment. First, it seems unnecessary to describe, as the author does, the construction and manipulation of universal stages which are little used now, such as the Leitz and especially the Winckel-Zeiss stages. On the other hand, it would have been advisable to give a diagram of corrections for the difference between the refractive index of the mineral and the hemispheres, rather than refer the reader to some other text which may not be immediately available. In discussing objectives used with microscopes of domestic construction, the author does not, for some reason, mention the x9 objective, although it is commonly used with the MIN microscope, especially in working with small grains. On p. 13 (§7), in describing the method of determining the vibration directions of the nicols (with biotite), the author omits a simpler and more convenient method of using partial polarization of light by reflection. It is not clear why. On p. 17 (§12), the author recommends adjusting the centering of the stage and thin section by rotation about the axis H (N-S). The non-coincidence of the plane of the thin section with the point of intersection of the stage axes is much more easily discovered by rotating the stage about the axis J (O. E-W) (by the movement of a point on the section along the vertical cross-hair).

The second chapter, "Determination of plagioclase on the Fedorov Stage," is the main part of the manual (46 pages). It gives a brief discussion of the indicatrix, the main methods of orienting the principal sections of the indicatrix on the Fedorov stage, and plotting of their positions on the Wulff net; and the methods of identifying axes of the indicatrix, the axial angle, and the optic character of the mineral. The author describes in considerable detail the twinning laws and optical properties of the plagioclase feldspars, the methods of finding the position of the twin axis on the Wulff net and directly on the stage, and the method of finding the twin axis coordinates and the pole of the composition face in the plagioclases. The concluding sections of the chapter present in condensed form, Fedorov's and Nikigin's twinning diagrams and the methods of determining the twinning law and the composition of the plagioclase by means of these diagrams.

The most important shortcomings of this chapter, in the reviewer's opinion, are as follows. First, Nikitin's twinning diagram (in one quadrant) is omitted and without it the worker using Fedorov's method cannot complete the determination of the plagioclase on the Fedorov stage. The author justifies this omission by saying that this diagram is given in the texts by Saranchina

[1]Gosgeoltekhizdat, 1957, 10,000 copies.

nd Soboleva, but this is not a valid excuse nd the recommendation that "for a com- lete mastery of the method only the circu- ar diagrams should be used" (p. 57) is sur- rising, since these diagrams are not gradu- ted (Figs. 24 and 25) and are therefore ractically useless.

Another serious shortcoming is the omis- ion of the zone method, whose advantages re mentioned on pp. 6 and 63. It should e noted that A. Rittman's book, "The Zone Method", (translated by Yu. A. Kuznetsova), as long since become a bibliographic rarity, nd that the accounts of this method in obolev's text and in Yeliseyev's "Methods f Petrographic Investigations" are insuffi- iently detailed. Therefore, a brief but omplete account of the zone method, which s commonly used in petrographic investiga- ions, especially in the determination of nicrolites, would have been very appro- riate in Dobrokhotova's manual. Unfortu- ately, this method is not given enough ttention.

The presentation of material on the etermination of orientation of the indicatrix nd identification of its axes cannot be con- idered successful. It is incorrect, methodo- ogically, to describe the methods of iden- ification of the indicatrix axes after plot- ing all of its planes and axes on the Vulff net. By doing this, the author makes he beginner perform unnecessary operations n the Fedorov stage. The determination of he indicatrix axes must go hand in hand vith the determination of the orientation of he ellipsoid, thus saving the worker's time, nd making the whole procedure more intel- igent. For this reason, the discussion of he "usual" method of identifying the indica- rix axes (as distinct from the rapid method, p. 37-42) is unjustified, in the opinion of he reviewer; it encumbers the manual with nnecessary material presented in incorrect equence.

On p. 23, the author states that in order o locate the optic plane, the microscope tage should be turned through 45°. The ase when axis β coincides with the axis $\underline{J}$ O. E-W) is discussed first, and the author vrites, "This means that the symmetry lane of the microscope coincides with the ptic plane of the indicatrix which contains oth optic axes and the axes β and λ. In this ase, in rotation about the axis $\underline{J}$ there nust come a moment when one of the optic xes will point at the observer's eye." But he microscope stage has already been urned through 45° and the optic plane no onger coincides with the symmetry plane f the microscope! Such inexactness in xposition can only confuse the beginner.

The precision of orientation of the prin- cipal section of the indicatrix and of the optic axes of the mineral is greatly in- creased if the extinction is checked with a compensator (quartz wedge), first during rotation about the $\underline{J}$ axis and then during rotation about the $\underline{A}$ ($\underline{M}$) axis. The author does not mention the use of a compensator in such cases.

On pp. 31-32, in dealing with the method of identification of the indicatrix axes by means of the Wulff net (hardly a convenient method), the author makes a very unclear statement to the effect that if the color of the mineral is increased on the introduction of a compensator, "the directions of the greatest and least refractive indices coin- cide in the compensator and the mineral." A similar statement which can be inter- preted in the sense opposite to that intended occurs further, in the second paragraph of p. 32. It is obvious that such formulations can only confuse the beginner.

It is an omission on the author's part that after characterizing all of the twinning laws in the plagioclase feldspars, she does not give the whole sequence of opera- tions used in identifying the plagioclases by twinning. An enumeration of these opera- tions is absolutely essential, because in the process of determination of a plagioclase after all the elements of the indicatrices of both individuals of the twin have been plotted on the Wulff net, the investigator, after orienting the composition plane, must imme- diately identify the twin law (normal, paral- lel, or complex), which is very important for the subsequent work on the stage and on the net. Nothing is said about selection of the plagioclase grain for this work, and yet it is on this that the successful and correct identification of the plagioclase on the Fedo- rov stage largely depends. In order to pre- vent possible errors, in determining the coordinates of the twin axis and the pole of the composition face, the author should have pointed out the permissible limit of deviation of these coordinates from the like indicatrix axes of the two individuals of the twin.

In the concluding part of the chapter it would have been well to have discussed the determination of the composition of the plagioclase with the aid of the twin law dia- grams in greater detail, especially for those cases where the diagram presents difficul- ties in the choice of the curve for this or that twin law (this happens commonly in determining sodic plagioclases). In particu- lar, it would have been desirable to list here the additional data which always make it possible to obtain correct solutions (ex- tinction angle sign, optic sign, and the axial angle, refractive indices, etc.).

Finally, it should be pointed out that in 19 (p. 58) the circular Nikitin-Fedorov diagram is incorrectly named polar net. This diagram is a combination of polar and two meridional nets crossing each other.

In the third chapter the method of determining alkalic feldspars on the Fedorov stage is given. It gives all that is necessary to acquaint the investigator with the principles involved and some specific points of the method. This chapter does not require criticism.

In the final, fourth chapter of the manual, the author describes the methods of determination of certain characteristics of colored minerals on the Fedorov stage, such as 2V, the angle between cleavage planes, the extinction angle, and the pleochroic formula. All these problems are discussed completely enough within the limits set by the author; however, in describing the determination of extinction angles in pyroxenes and amphiboles, the use of twinning seams should have been mentioned as an occasional substitute for cleavage traces. The method worked out by D. S. Korzhinskiy is not mentioned, and the term "adsorption" is used incorrectly as applied to the pleochroism of colored minerals; the phenomenon of pleochroism is related not to adsorption of light but to its adsorption in passing through a mineral.

It is difficult to agree with the author's skeptical attitude towards the precision in determination of the thickness of a thin section on the Fedorov stage and with the statement that for this reason $\lambda - \alpha$ is no longer determined on the stage (p. 78). Undoubtedly the insufficient exactness in the determination of the thickness of thin sections by various methods is at present a "bottleneck" in the determination of the birefringence of minerals on the Fedorov stage. Still, such determinations are much more exact than the visual estimates used in petrography. Therefore, the pessimistic conclusion of the author appears unfounded.

In conclusion it may be said that Dobrokhotova's manual on the Fedorov method, in general useful and welcome, is not free from shortcomings, mainly of methodological character. It is hoped that the author will consider these critical remarks and introduce appropriate corrections into the second edition of her manual.

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CHRONICLE

ACTIVITIES OF THE INTERNATIONAL VOLCANOLOGICAL ASSOCIATION

At the Eleventh General Assembly of the International Union of Geodesy and Geophysics, the International Volcanological Association (IVA) made a resolution to broaden its field of activity.

In connection with this resolution, the Central Bureau of the Association was reorganized. Instead of the single section on paleovolcanology formerly included in the IVA, the following four sections were organized: 1) Section of Active Volcanism, 2) Section of Volcanophysics, 3) Section of Physical Chemistry of Magmas and 4) Section of Paleovolcanology and Plutonism, presided over, respectively, by M. Neumann van Padang (Netherlands), G. S. Gorshkov (U. S. S. R), H. Kuno (Japan) and B. Geze (France). A. Rittmann (Switzerland) was reelected president of the International Volcanological Association, and F. Signore (Italy), the general secretary.

The New Central Bureau of the IVA made plans for future research, which are of interest not only to volcanologists but also to geologists in general and especially to petrologists. The content of the projected research for each section is as follows:

1. The Section of Active Volcanoes will stress the study of methods of protection of men and property against volcanic eruptions and organize a symposium on "Prediction of time, place, and character of eruptions, the direction of lava flows and nuee ardentes and methods of controlling them." It will once again request the National Committees of the countries with active volcanoes to report volcanic episodes as soon as possible. In cooperation with the section on the Physical Chemistry of Magmas, it will increase work on collection and study of volcanic gases and continue the "Catalogue of the Active Volcanoes of the World," since it includes information of great importance for further research in volcanology and the allied fields of geophysics.

2. The Section of Volcanophysics will broaden seismometric, thermal, gravimetric, and other investigations of active volcanoes and conduct symposia at the coming Twelfth General Assembly of 1960 in Helsinki on the following subjects: 1) the application of geophysical methods to volcanology, 2) the use of volcanic energy for practical and scientific purposes.

Besides this, the Section expressed a desire for a joint symposium with the International Seismological Association and Physics of the Earth's Interior on the "Relationship between magmatic and geodynamical phenomena," also to be held in Helsinki.

3. The Section of Physical Chemistry of Magmas. The Section will conduct spectographic studies of volcanic gases directly at the active volcanoes, work on the development of apparatus and methods for collecting magmatic gases for a thorough investigation including that of their isotopic composition and their content of dispersed elements. It will investigate the equilibria of the liquid, solid, and gaseous phases in magmas, the process of their formation in nature during volcanic activity, and the distribution of dispersed elements in all three magmatic phases. It will make rock collections, even small ones, with precise indication of locality and the nature of structure and stratigraphy of the volcano. It will collect complete chemical analyses of newly erupted rocks.

4. The Section of Paleovolcanism and Plutonism. The Section will conduct investigations on the relation between volcanoes and plutons and prepare a symposium on this problem based on well known regional examples, and will prepare a symposium on ignimbrites, their origin, and the sequence in the variation of lavas in different regions.

V. I. Vlodavets